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The Problem of Subdivision of Metamorphic Rock Sequences for Geological Mapping

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Abstract—Problems concerning the methodology of subdivision and mapping of metamorphic rock units, their taxonomy, and nomenclature are discussed. Several of the most important concepts are specified, and the necessity of elaborating the taxonomic system of metamorphic units and their nomenclature is substantiated. Regional subdivisions, such as metamorphic rock complexes, are considered to represent taxons of the highest hierarchic rank because of their significant role in the structure of major elements of the earth crust, such as structural–formational zones (regions). The main mappable units are taxons of a lower rank identified as metamorphic rock formations (stable parageneses of metamorphic rocks). Metamorphic units represent secondary geological bodies and should not be considered and mapped in general (not specialized) geological maps as units equal in rank to primary (stratigraphic and plutonic) units.

Key words: *metamorphic rock complexes, metamorphic formations, taxonomy, mapping, petrographic code.*

Individualization of every cartographic geological body in terms of spatial–temporal relationships is a main peculiarity of the geologic–cartographic method. Precisely the individualization of geological bodies is an aim of the procedure of subdivision and correlation when studying sedimentary, magmatic, and metamorphic rocks. Elements of a stratigraphic scale, i.e., stratigraphic units (environmental records) and boundaries (event records), are chronologically unrepeatable and unique. This is the basis of all stratigraphic constructions (Meyen, 1989). Intrusive rocks, which are distinguished as a special type of lithodemic or lithostratigraphic units in some stratigraphic codes (*North American ...*, 1983; *International Subcommission ...*, 1992), also possess (in a broad sense) a definite status in stratigraphy.

The same stratigraphic approach was used by geologists in their attempts to subdivide the intensively altered metamorphic rock complexes. Their stratiform (or better to say stratumlike) structure was interpreted as a normal bedding and imparted with properties of the stratigraphic sequence (succession). However, this approach in general has not proved its worth (Shul'diner, 1992). As for the cartographic stratumlike elements of the metamorphic rock complexes (metamorphic lithostratigraphic units), it is impossible, when they are strongly altered, either to establish their primary nature, or to define the primary events reflected by boundaries between them. The monogenetic origin of rock bodies subsequently metamorphosed requires a special substantiation, which is usually insufficient despite their classification as para- or orthorocks. The stratumlike structure of particular metamorphic rock bodies can mask various primary textures and geological features, such as isoclinal folding, disjunctive dislocations, zones of parallel magmatic dikes, stratigraphic unconformities, zones of tectonic disintegration, and metamorphic or metasomatic alterations. The geology of crystalline metamorphic rock complexes lacks indicators of temporal succession reflected in boundaries between rock bodies of different age that are characteristic of sedimentary and magmatic deposits. Unlike the latter rock complexes, metamorphic bodies are characterized by the prevalence of structural and petrographic features that resulted from the alteration of the features the reflecting conditions of their origin. Subdivision and mapping of metamorphic rock complexes do not reveal the chronological individuality of their constituent elements and do not make the status of metamorphic units of the lower rank definite in stratigraphy. On the whole, the results of a geologic investigation of a crystalline metamorphic rock complex reflect only the first, structural–lithological stage of subdivision and correlation.

Realization of this situation resulted in the rejection of the stratigraphic approach and in the subdivision of metamorphic rock complexes on the basis of their structural and petrographic properties, as is reflected in recent editions of the petrographic code (*Petrograficheskii ...*, 1995) and manuals for geologic map compilation (*Instruktsiya ...*, 1995).

The Petrographic code states: "Metamorphic complexes of rocks, which significantly or completely lost primary features during metamorphism so that their origin cannot be reconstructed using direct external features, can be defined as independent petrographic

units .... Distinguishing and cartographic works on metamorphic rock complexes should be performed using real observable structural and petrographic features. A geologic map should demonstrate metamorphic rock complexes defined according to these features .... Reconstruction of a protolith is allowed in this case only as interpretation and should not be shown in the map" (*Petrograficheskii ...*, 1995, p. 47). The cited statement means (although it is not clearly formulated) that metamorphic rock complexes should be represented in geologic maps as independent units having equal rights with stratigraphic and plutonic units. This innovation appears to be rather surprising. Practice of structural and petrographic subdivision of metamorphic rocks is scarce, and its methodology is poorly elaborated. The petrographic code that describes in detail principles of subdivision, correlation, and description of magmatic units gives only schematic, insufficient for mapping, characteristics of the same principles relative to metamorphic objects. The authors of the recent version of the code cannot be blamed for this deficiency because it reflects the real state of the considered problem. Nevertheless, compilation of geologic maps of a new generation (scales 1 : 100 000 and 1 : 200 000) is already in progress, and these methodological shortcomings should inevitably affect the map quality.

The considered problem is not simple and requires further discussion. Here, we will consider three main aspects concerning the methodology and specifics of distinguishing and mapping metamorphic rock units, their taxonomy, and nomenclature.

1. *The methodological base of defining the metamorphic rock units in geologic maps.* It is a common speculation (and this comes out in one way or another from statements of various methodological handbooks) that since metamorphic rocks represent, like sedimentary and effusive rocks, an independent petrographic class, geologic units formed by these rocks should have a status in geological maps equal to that of stratigraphic and plutonic units. Meanwhile, as is known, classification of cartographic geological units is by no means based on petrographic features and does not correspond to the petrographic classification of rocks. Geological bodies, not rocks, are mapped, and they are classified according to the mode and time of formation in geological space, not according to their composition. From this point of view, the cartographic geological units are divided into two geologic–cartographic classes, namely, into primary and secondary units. The primary units comprise geological bodies whose successive generation resulted in the formation of new portions of the earth crust. Their further classification is based on two features: origin (mode of formation), by which they are divided into stratigraphic and plutonic units (which, we repeat, does not correspond to subdivision of rocks into petrographic types and classes), and age. The main illustrative principle of a geological map, the color background (or hatching in black and white maps), is used for designation of these two features. Because the initial form and occurrence patterns of stratigraphic units are known a priori and for plutons they can be inferred from their relationships with stratigraphic units, representation of these units on a geological map gives an adequate idea of the mode and formation time of the pertinent earth crust segment. Combination of these initial characteristics with the recent situation represented in the map allows the structural evolution of the considered earth crust fragment to be interpreted.

The class of secondary units includes geological bodies that were formed as a result of alteration of existing geological bodies during structural and compositional transformations of their rocks. These are mainly metamorphic geological bodies—aureoles and zones of geothermal, dislocation, and other types of metamorphism. However, certain sedimentary bodies (for instance, weathering crusts located in situ) are included in this class, so the disagreement with petrographic classification is quite evident. Principal requirements of the standard geologic–cartographic model, such as primary genetic and chronological individuality of geological bodies, are not applicable to secondary units. The fundamental difference between secondary and primary units is expressed in the map through the pertinent symbols. Less expressive graphic means like various specks and hatching superimposed on the color background of primary units are usually used to show secondary units. Often, they are without graphic representation and supplemented only by written characteristics of secondary alterations in the legend to the map. The possibilities of graphic means are not unlimited, and the aim of the general (not specialized) geologic map when using these possibilities is to present spatial and temporal relationships between primary geological objects, whereas the information on secondary units can be limited by a certain minimum of information determined by their significance in particular cases.

In other words, the preeminence of primary objects over secondary ones is a peculiar feature of a general geologic map. No matter how great the significance of metamorphic units may be in one or another situation, they are always of a secondary nature and cannot be considered as having equal rights with the objects of the primary class—stratigraphic and plutonic units.

2. *Cartographic specifics of strongly metamorphosed rocks.* In the Petrographic Code, it is suggested to discriminate, depending on the degree of rock alteration, between metamorphic rocks (complexes) proper and metamorphosed rocks (complexes). Criteria for their discrimination are formulated unclearly. Metamorphic rocks are products of deep transformation and elimination of "initial features of structure, texture, and mineral composition." Metamorphosed rocks are "the rocks that preserve, to one degree or another, some of their primary features, which help to define both their origin ... and initial petrographic composition" (*Petro-*

graficheskii ..., 1995, p. 26). These two rock types, one with features sufficiently preserved to determine reliably the protolith and another lacking initial properties, are divided by a wide gap filled with intermediate varieties. Another chapter of the code (p. 47) contains a more exact definition, according to which the interpretation of substratum and, correspondingly, subdivision into metamorphic and metamorphosed rocks should be performed only by using "observable features" directly during field investigations.

In this context, contraposition of the terms "metamorphosed" and "metamorphic" is etymologically unfortunate and also bears an element of indefiniteness: neither of them indicates the degree of alteration, stating only the fact and the result of metamorphism.

In fact, the requirement of dividing metamorphic rocks according to the degree of preservation of their initial properties is narrow and should not be legalized by the code. It is evident that identification of protoliths depends not only on the degree of insight into the metamorphic rock properties but also on the possibilities of the methods used, which are constantly being improved. Therefore, universal criteria for rock discrimination according to the above feature are absent and should not be regulated. The requirements of geologic mapping oblige geologists to gain maximum insight into the premetamorphic history of a substratum, and this goal should not be restricted by any preset limit (including the possibilities of the field identification). The studied unit should be classified as either premetamorphic (i.e., stratigraphic or plutonic) or metamorphic, depending on the particular result obtained but not on the object properties.

The problem of mapping metamorphic rock units, i.e., geological bodies composed of strongly metamorphosed rocks with unrecognizable substratum, is that only one of the three characteristics required for mapping—the spatial parameters—can be reliably established, whereas the initial origin and successive stages of the formation commonly remain undefined. When the above-discussed geologic–cartographic principles are strictly followed, these bodies should be designated in maps using the "metamorphic specks" against a white background or color age background of a certain undivided stratigraphic unit.

However, when metamorphic rock units with the undivided primary substratum occupy large areas in the map, such a method of presentation is no longer rational, because the main descriptive medium—color—appears to be used inefficiently. In this case, the application of a color background seems admissible to designate high-temperature metamorphic rock units that lost reliable indicators of their initial nature. However, it would be wrong to think that, using this procedure, we level their status with that of primary rock units. Metamorphic rock units are of dual nature and, in the considered case, represent the undivided primary (stratigraphic and plutonic) rock bodies unrecognizable due to superimposed alterations. The necessity of defining metamorphic rock units in the map as independent elements is determined by our insufficient knowledge and the impossibility of reliable substantiation of initial units rather than by their metamorphic nature. As cartographic units, they have a temporal status, and as they come to be better understood, primary bodies will be distinguished within them and the method of their designation will be changed.[1]

The selection of illustrative means to map metamorphic rock complexes depends on the degree of alteration, as well as on the map scale and its possible resolution. The necessity in background representation first appears in the compilation of large-scale maps of areas with widespread metamorphic rocks of indefinite primary nature. In this case, the parallel application of color for both primary and secondary units appears to be reasonable because of the necessity to show the factual basis of geological synthesis. In the small-scale maps, it is preferable to follow the uniform principles of representation using color only for primary units.

The system of age indices of metamorphic rock units in geological maps should be consistent with the suggested approach and required uniformity of geological record: the age indices should denote the established or suggested time of rock formation, but not alteration. The position of metamorphic rock complexes in the legend and stratigraphic column should also correspond to the substratum age. Infringement of this rule will result in awkward cartography: units of the same metamorphic complex with similar substratum may occur in the legend at different age levels, if its weakly metamorphosed zones are classified according to the substratum age, and strongly metamorphosed rocks, according to the age of metamorphism. It may happen that intrusions will be shown to intrude metamorphic supracrustal rocks, which, according to their age index, may turn out to be younger than intrusions, and so on. In most of the handbooks, the rules of age designation of metamorphic rock units are not specified.

3. *Taxonomy and nomenclature of metamorphic rock units.* According to the Petrographic Code, the metamorphic rock complex represents the main metamorphic unit. However, the concept "metamorphic complex" and its scope are outlined indistinctly. The definition is as follows: a metamorphic complex is "an association of metamorphic rocks composing geological bodies that were formed in a certain geological space and time and bear stable features of composition, structure, texture, and relationship with the surroundings" (*Petrograficheskii* ..., 1995, p. 46). The fact that complexes were formed "in a certain geological space and time" is indisputable; however, the statement that

[1] In using this approach, the main difficulty is in the evaluation of reconstruction reliability, because this category is flexible and depends on the general level of knowledge. However, this is a problem that cannot be regulated, and its solution in every particular case is the responsibility of investigators and controlling institutions.

they bear stable features of composition, structure, and texture, on the contrary, is invalid. And what does "relationship with surroundings" mean?

As a whole, this definition is a copy of the magmatic rock complex definition given in the same code. However, in the latter, the authors specify the scope of the complex, which is determined by its formation "in a certain geologic setting during the evolution of a single ... intrusive process" (p. 41). In the case of the metamorphic complex, any explanation of this kind is missing, and the question on the limits of the "metamorphic complex" remains without an answer. The similarity of wordings does not help, because the different origin of magmatic and metamorphic rocks means a principal difference between properties used to unite two rock types into a complex, and, thus, there is no correspondence between rock units formally of the same rank (Shul'diner and Ermolaev, 1989). The interpretation of magmatic and metamorphic rock complexes in the Petrographic Code to the extent that they are identical in the aspect of cartography and formational analysis seems to be incorrect.

Indeed, it is difficult the define the concept of the "metamorphic rock complex" because of its multiple meanings. However, it is possible to explain the meaning of this term as it is accepted in publications. First, this unit is of a high hierarchic rank. A metamorphic rock complex is the result of a regional metamorphism and, correspondingly, represents a regional unit, as is also noted in the code. With respect to its extent and significance, the concept of metamorphic complex can be correlated with the concept of a structural–formational zone, or a region. As a rule, dimensions of metamorphic rock complexes outcropping at the surface are quite large and, by the definition of V.A. Glebovitskii, commensurable with dimensions of structural–formational zones. The metamorphic rock complex characterizes the essential peculiarities of the structural–formational zone where it is exposed, and, in many respects, it determines its specifics. Petrographic properties of the metamorphic rock complex are not necessary stable, but their combination allows the complex to be characterized as a genetically integral geological body (group of bodies) and its (their) place in the geological structure and regional geological history to be outlined.[2]

[2] In this connection, the necessity of defining petrotypes of metamorphic rock complexes seems doubtful. The metamorphic rock complex is a complicated large geological construction, which can be considered as an integral object, and therefore any of its part, or any separate outcrop selected to represent its petrotype can impoverish and distort its content. It appears that, to be representative, the petrotype should be equal to the entire unit. Therefore, the code requirement relative to metamorphic rocks (p. 53) to select and describe by all means their petrotype should be abolished. Simultaneously, it seems to be appropriate to supplement characteristics of any petrographic unit (including those of a lower rank) with thorough description of the most well-studied reference objects, i.e., the standards whose function, as is known (Meyen, 1989), differs from that of a holotype.

Such an understanding of the "metamorphic complex" is popular in most works on metamorphic rocks since the first works by D.S. Korzhinskii (1939 and others), where he defined the Aldan, Drevnestanovoi, and Slyudyanka rock complexes of East Siberia; these rock complexes retain their significance as major regional metamorphic units even today.

The problem of subdivision of metamorphic rock complexes into smaller units is considered neither in the Petrographic Code nor in handbooks.

Meanwhile, because of their large dimensions, metamorphic rock complexes, i.e., petrographic units of the highest rank, generally cannot serve as elementary cartographic units. Cartographic practice shows that in the geologic maps of scales 1 : 50000, 1 : 200000, and, often, 1 : 1 000000, petrographically uniform bodies composed of one–three and, sometimes, more types of metamorphic rocks are commonly defined. These bodies are mostly of stratal or lenticular form, and, being several kilometers thick, they can be traced over many kilometers or even tens of kilometers along the strike. The units of this rank are shown in many geological maps of this country (true, often without proper substantiation) as stratigraphic units of the formation or group rank. In publications dealing with formation analysis and mapping, units of this rank are named as metamorphic formations, paragenerations, and petroformations (*Metodicheskie* ..., 1970; Travin, 1975; Shul'diner, 1992).

In such a scope, the metamorphic formation is the most usable taxon of metamorphic units, usually satisfying the subdivision requirements. Sometimes, there is a necessity to unite formations. In this case, the taxon of an intermediate rank is recommended to be the metamorphic group (paragenetic formations). Auxiliary taxonomic units, such as subcomplex, subgroup, and subformation (Shul'diner, 1992), can also be used. Occurrences of local (contact, dislocation) metamorphism are mostly referred to units of the lowest rank.

In connection with the application of the term "metamorphic formation," let us discuss another statement of the Petrographic Code: "Mapping of geological rock bodies and subsequent formation analysis of their combinations united into magmatic (metamorphic) complexes represent independent and unequal elements of regional geologic investigations. Mapping and geological–petrographic study of a certain rock complex is not always accompanied by the subsequent formation analysis; however, this kind of work should always precede the latter .... It is incorrect, methodologically and in essence, to raise the question of direct geological mapping of magmatic and metamorphic formations" (*Petrograficheskii* ..., 1995, p. 16). This statement appears to be valid only in that narrow meaning for which it is formulated above, i.e., relative to petrographic units of high taxonomic rank. In a more general sense, it cannot be accepted for several reasons.

First, in the formation analysis, the term "formation" is traditionally used for two ideas: for designation of the individual geologic body and for designation of the classification group (type, class) to which this body belongs. This duality was noted by Yu.A. Kuznetsov, who wrote about "concrete" and "abstract" formations. In magmatic geology, the particular formation is designated by the term "complex," whereas the magmatic formation bears its abstract meaning. However, the geology dealing with sedimentary and metamorphic units lacks special designations, and the term "formation" is used in its both meanings. The ambiguity of this term is aggravated by the fact that it is used for designation of subdivisions of a different taxonomic rank. This situation creates a certain inconvenience; however, it cannot be improved through the prohibiting decree; it is necessary first to improve the nomenclature system.

Here, we should again emphasize significant differences in application of the term "formation" relative to magmatic and metamorphic rocks. The magmatic formation has both genetic and hierarchic definiteness. Introduction of its dual (abstract and concrete) meaning favors insight into the regularities of the evolution of magmatism in time and space. The metamorphic formation is a particular structural–petrographic and cartographic unit whose primary genetic individuality is assumed rather than proven. With the current state of knowledge, metamorphic formations remain in the status of an empirical generalization. Hierarchically (not genetically), they may correspond to stratigraphic units and plutons of various ranks or to their combinations. The place of metamorphic formations in the hierarchy of metamorphic rock units is defined approximately. Simultaneously, the metamorphic formation, with due regard to the limited possibility of its chronological individualization, appears as an abstract formation, i.e., a sort of reference lithologic unit.

Another aspect of the problem is the following. Every geological body, from a crystal to the largest rock complex, is an individual object and simultaneously a representative of its type or class. In this case, the object individuality is determined by two factors—internal and external. The former is responsible for specific features of the object, its originality, and uniqueness, and the latter, for its peculiar position relative to other objects of the same rank in their common structure. The higher the hierarchic rank of the unit, the more complicated its structure and the stronger its specifics, i.e., the internal factor of individuality. That is what the metamorphic rock complexes are—the complicated, internally organized bodies, whose unique individual features often "screen" their general formational properties. Identification of these latter becomes, as is stressed in the code, a matter of special analysis separated from direct problems of geological mapping.

In the formational units of the lower rank, the internal specifics is commonly expressed less distinctly, and their individuality is determined by the external factor. For instance, there are innumerable, difficultly distinguished sandy–silty sequences, every one of which, however, is characterized by a strictly definite position in the stratigraphic sections, which makes it represent an individual and chronologically unique unit, e.g., formation N, horizon M, and so on. Similarly, if a granite lacking individual features forms one of the phases or facies of a given, chronologically individualized plutonic complex P, the individuality of this granite is granted. Owing to the lack of clear internal specifics, units of the lower rank serve as carriers of type features, i.e., as representatives of "abstract" formations, which can be easily identified during the geological mapping. Moreover, units of such a rank are defined, as a rule, primarily according to type features and only later acquire their individual status. In the case of metamorphic low-ranked units, this second procedure—identification of an individual (individualization)—may fail when structural, age, and genetic relationships of metamorphic bodies with each other and, correspondingly, their place in the metamorphic complex turn out to be undeterminable.

Thus, when defining metamorphic rock units of the lower rank, the topic of prime importance is their type characteristics displaying the metamorphic rock affinity to a certain recurrent paragenesis, i.e., to the metamorphic formation. In this case, it is reasonable to speak about direct mapping of metamorphic formations. The application of the term "formation" still remains ambiguous; however, the lack of developed nomenclature compels us to reconcile with this shortcoming. Anyway, the meaning of this term is clear as a rule from the text.

The aforesaid suggests certain rules of terming the cartographic metamorphic rock units. According to the Petrographic Code, the "petrographic name of the complex should be derived from names of prevailing or most characteristic ... rocks composing it" (*Petrograficheskii* ..., 1995, p. 52). In fact, this rule is applied only to metamorphic units of the lowest rank—to metamorphic formations (for instance, hypersthene–gneiss, amphibolite, or quartzite–kinzigite formations, etc.). However, often, it is impossible to characterize in a similar way an integral but compositionally variable unit such as a metamorphic rock complex. Petrographic names of these latter should reflect the most general properties of particular complexes outlining their specifics such as the zoning, degree of metamorphism, depth of origin, and other features. The method of selecting these features in a given case cannot be standardized (for instance, the Aldan granulite complex, the Lapland complex of high-pressure granulites, the Ladoga zonal low-pressure complex, the Belomorian polymetamorphic high-pressure complex, and so on). At the same time, precisely the geographic name of the metamorphic complex, with due regard to the specifics of this unit, serves as a main "identification label." On the contrary, the particular metamorphic formation is labeled by its composition and relation to the particular

metamorphic complex; therefore, the proper geographic name is excessive for this unit. The sole exception are those formations whose features are proven to be unique, individual.

CONCLUSION

Problems discussed in this paper in connection with the subdivision of metamorphic rocks for the cartographic purpose concern the refinement of concepts and some ideas on a particular category of cartographic units, such as metamorphic rock complexes and formations. Since a geological map represents a multifactor model of geological reality that discloses modes and succession of formation and transformation of geological bodies, the graphic means of map compilation are and should be differentiated in respect to primary and secondary (superimposed) geological units. Metamorphic rock units are of the secondary type, and, in general geological maps, they can be represented as independent cartographic units instead of primary units only in those cases when they can be defined and subdivided only on the basis of their structural–petrographic properties, in view of impossibility of identifying their primary nature and geological age. Moreover, metamorphic rock units should not be considered as identical in status to the primary stratigraphic and plutonic units; they represent only a result of the initial stage of subdivision and correlation, whose final aim is the chronological individualization and identification of primary units of the geologic–cartographic class.

The highest taxon of metamorphic rock units is represented by regional metamorphic complexes, whose hierarchic rank reflects their significant role in the structure of major elements of the earth crust, such as structural–formational zones (regions). Taxons of the lower rank—metamorphic formations (stable recurrent parageneses of metamorphic rocks)—are the main cartographic units. Individual (geographic) and type (formational) names of metamorphic units should reflect real relationships between individual and type properties of various taxons.

Reviewer M.A. Semikhatov

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The Ordovician Ostracods and Articulate Brachiopods from the Gavrilov Yam Well Drilled in the Moscow Syneclise

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Abstract—The results of a comprehensive study of the Ordovician lithology, ostracods, and articulate brachiopods recovered from the Ordovician part of the sections of Gavrilov Yam Wells 1 and 5 are presented. The subdivision of the Ordovician sequence is based on the regional stratigraphic scheme of the East European platform (*Resolutions ...*, 1992). The identified ostracod assemblage contains more than 50 species irregularly distributed throughout the section. A diverse ostracod assemblage (about 65% of the total diversity) was recovered from the Middle Ordovician deposits, but their finds in the Lower Ordovician beds were scanty. The least diverse ostracods were found in the Upper Ordovician part of the section. An analysis of the species diversity showed numerous Middle Ordovician ostracod species to be common for both the Moscow syneclise and Baltoscandia. This feature implies a close relation between these basins. Six new ostracod species are described belonging to six genera, two of which are also new. The most typical Ordovician ostracod species of the Moscow syneclise are pictured for the first time. In the studied sections, the remains of articulate brachiopods of 17 species were recovered, with the exception of one, from the carbonate deposits of the Lower and Middle Ordovician. The Upper Ordovician deposits have almost no articulate brachiopods, apparently owing to their specific, perhaps non-marine genesis.

Key words: *ostracods, articulate brachiopods, biostratigraphy, lithology, Ordovician, Moscow syneclise.*

INTRODUCTION

The Ordovician deposits of the Moscow syneclise contain different faunal groups: inarticulate and articulate brachiopods, ostracods, trilobites, conodonts, echinoderms, bryozoans, and graptolites. Information on ostracods from the Moscow syneclise is scanty (Dmitrovskaya, 1989a; Prokof'ev and Kuznetsov, 1982). In most cases, only some ostracod genera were identified, and before now neither their description nor their pictures had been published. The articulate brachiopods, though studied in greater detail (Alikhova, 1969; Dmitrovskaya, 1989b), still require a more accurate insight into their distribution in the Ordovician rocks recovered from the deep wells drilled in the Moscow syneclise, because they were previously studied mainly in natural exposures in the northeastern part of the East European platform.

In view of this and also to elucidate the stratigraphic significance of these faunal groups, we sampled sections of two reference wells Gavrilov Yam 1 and 5. Both wells were drilled south of Yaroslavl in the Gavrilov Yam geophysical test area located in the southern structural zone on the Rostov swell, which was determined by geophysical data (Dmitrovskaya, 1991). The Ordovician sequence is given according to the regional stratigraphic scheme for the central areas of the East European platform, adopted in November 1991 (*Resolutions ...*, 1992) (Fig. 1).

The 85 samples with ostracods from the Gavrilov Yam section were analyzed. Initially, the entire Ordovician part of the cores obtained from both wells was studied, then only those rock samples that visually revealed ostracod remains were selected. The sampled rocks were mainly organogenic and clayey limestones and shales. In order to extract the ostracod remains from more solid rocks, the usual procedure was applied; i.e., the sample was crushed into fine grains, calcinated in a muffle kiln, and abruptly chilled. The substantially clayey rocks were alloyed with sodium hyposulfite or boiled in hydrogen peroxide.

The lithology of the sections from the Gavrilov Yam Wells 1 and 5 containing articulate brachiopod assemblages was studied and described by Yu.E. Dmitrovskaya; the ostracods were described by L.M. Mel'nikova, who also identified their new taxa and pictured the most characteristic species shown in Plates 1 and 2. The ostracod collection no. 4475 is deposited at the Paleontological Institute, Russian Academy of Sciences.

DESCRIPTION OF THE GAVRILOV YAM WELLS 1 AND 5

In Gavrilov Yam Well 1, the Ordovician deposits correspond to the depth intervals of 1500–1668 m, and

in Well 5, to that of 1495–1655 m. These deposits have the following peculiar features: (1) the argillaceous composition of the upper Bugino Subformation, which is a part of the It' Horizon; (2) the carbonate composition and reduced thickness of formations referred to the Volkhov (Nefedovo Formation), Kunda (Veksino Formation), Polomet' (Lunka Formation) and Berezaika (Skalino Formation) horizons; (3) in contrast to the sections in the central zone of the Moscow syneclise, our locality revealed a more complete stratigraphic range of the Kubeno Formation, owing to less intense erosion in the pre-Mitino time; (4) we also established the absence of upper Tremadocian (Ukhra Formation) and lower Arenigian (Sementsovo Formation) deposits. The

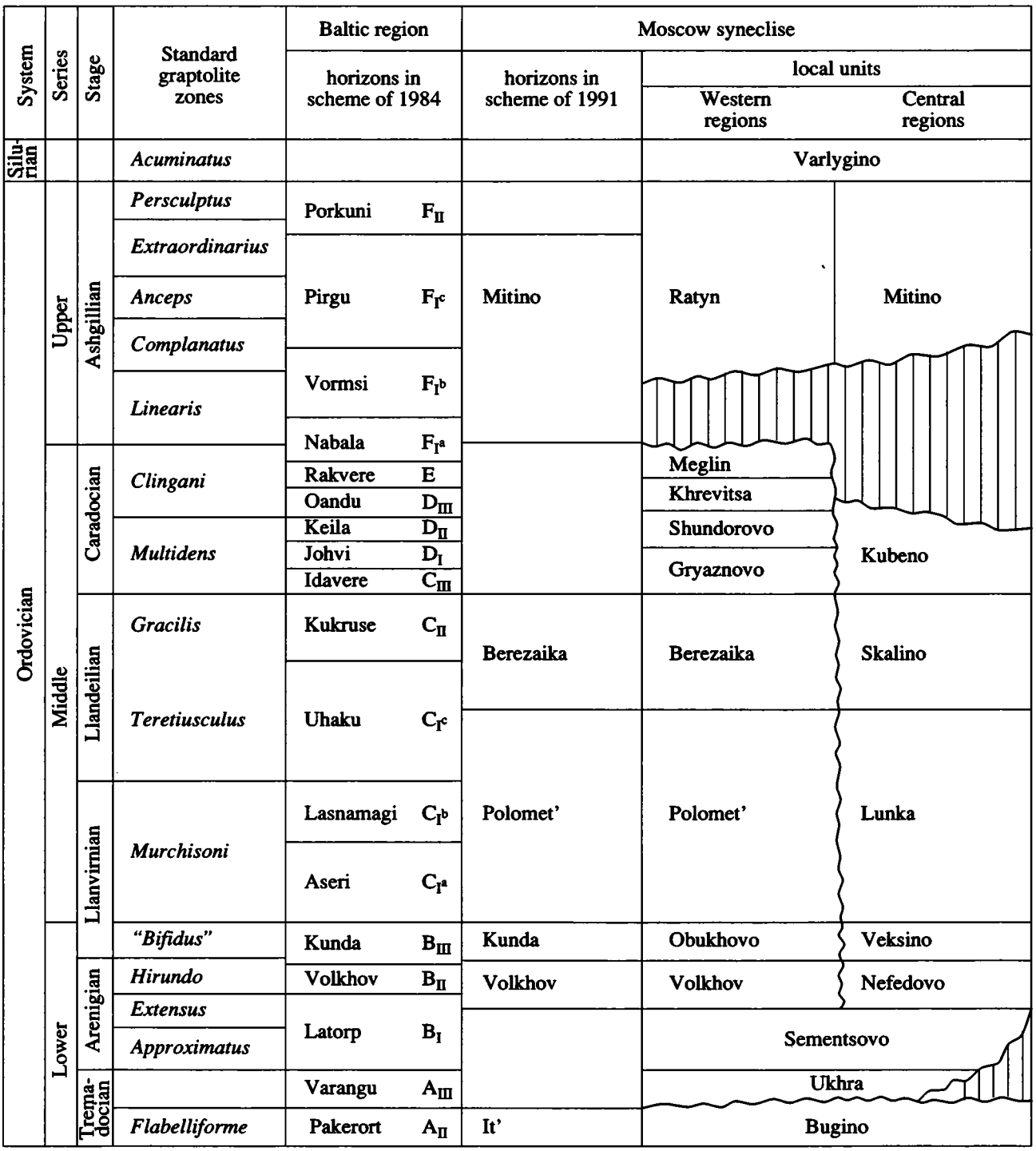


Fig. 1. Correlation of Ordovician deposits of the western and central regions of the East European platform (after Dmitrovskaya, 1991).

Ordovician deposits overlie the Upper Cambrian terrigenous rocks and are overlain by rocks of the Varlygino Formation (the Silurian), which are dolomitized to a certain extent.

Deposits in the sections of the Gavrilov Yam Wells 1 and 5 span three Ordovician subsystems, starting with the Lower Ordovician *upper Bugino Subformation* (It' Horizon; core recovery of 60%) at depth intervals of 1668–1640 m (Well 1) and 1655–1635 m (Well 5). The subformation is composed of greenish to dark gray carbonate-free shales and siltstones with thin beds of clayey fine-grained sandstones. The rocks contain only inarticulate brachiopods of the Lingulida order and acritarchs; these finds allowed us to relate the host rocks to the It' Horizon of the lower Tremadocian Substage. The ostracod remains were not observed in the upper Bugino Subformation.

The Nefedovo Formation (Volkhov Horizon) was reached at depth intervals of 1640–1628 m (Well 1; core recovery of 94%) and 1635–1623 m (Well 5). The formation is represented by gray and dark gray organogenic clayey limestones and greenish to ashy gray carbonate clays and marls with fine limestone beds and concretions. The following brachiopods were identified in this unit: *Paurorthis parva* (Pand.), *Productorthis* ex gr. *obtusa* (Pand.), *Ahtiella jentzschi* (Gag.), *Ingria nefedkewi* (Eichw.), and *Ingria* sp. nov. 1. These species allowed us to regard this formation as part of the Volkhov Horizon and to date it back to the Arenigian Stage.

The ostracod finds include *Conchoprimitia glauconitica* Opic, *C. gammae* Opik, *C. distincts* Sarv, *Piretopsis* (Protallinnella) *grewingkii* (Bock), *Piretopsis* (Pr.) *gavrilovensis* sp. nov., *Ogmoopsis veperi* Sarv, *Og. ramosa* Sarv, *Glossomorphites grandispinosus* (Hessland), *Asteusloffia acuta* (Krause), and *Tvaerenella* aff. *modesta* Sarv.

The Veksino Formation (Kunda Horizon) was reached at depth intervals of 1628–1624 m (Well 1; core recovery of 100%) and 1623–1610 m (Well 5). The lower part of the section is composed of greenish to dark gray, somewhat clayey limestones overlain by dark gray marls. At the base of the formation (depth interval of 1628–1626 m), *Ingria nefedjewi* (Eichw.) and *Ingria* sp. nov. 1 from the Nefedovo Formation can still be found. Higher up the section, the presence of *Athielle gemella* (Eichw.), *Ingria* sp. nov. 2, *Iru concava* (Pahl.), *Orthombonites* ex gr. *calligramma* (Pahl.), and *Antigonambonites* cf. *aequistriatus* (Gag.) allowed us to refer this formation to the Kunda Horizon of the upper Arenigian–lower Llanvirnian.

Among ostracods, there are also species similar to those from deposits of the Nefedovo Formation.

The Lunka Formation (Polomet' Horizon), recovered from depth intervals of 1614–1592 m (Well 1; core recovery of 56%) and 1610–1585 m (Well 5), is represented by gray to dark gray pelitomorphic limestones and dark gray marls with small pockets, interbeds, and lenses of organogenic–clastic and detrital limestones. The articulate brachiopod assemblage contains *Plectambonites planissimus* (Pand.), *Clitambonites adscendens* (Pand.), *Leptestia* (*Leptestia*) *humboldti* (Vern.), *Leptaena* (*Leptaena*) sp. nov., and *Sowerbyella* (*Viruella*) *orechovensis* Nik. These brachiopods refer the Lunka Formation to the Polomet' Horizon of the upper Llanvirnian and lower Llandeilian.

The ostracod assemblage of the Lunka Formation is diverse in composition and abundant. About 40 species were identified in this assemblage, including *Conchoprimitia gammae* Opik, *C. micropunctata* Hessland, *Nida nida* Sidarav., *Sigmoopsis cornuta* (Krause), *S. platyceras* (Opik), *S. lamina* Sarv, *Ctenonotella aculeata* sp. nov., *Longiscula arcuaris* Neckaja, *L. parrectis* Netckaja, and others (Fig. 2). Only ostracods of genera *Conchoprimitia, Tvaerenella,* and *Steusloffia,* though represented by other species, are inherited here from the lower assemblage.

The Skalino Formation (Berezaika Horizon) is located from depth intervals of 1592–1570 m (Well 1; core recovery of 75%) and 1585–1568 m (Well 5). It is composed of gray and bluish to ashy gray horizontally bedded marls, which are mostly dense, hard, and contain small lenses and bands of gray limestones 1.5–2.5 cm thick. The presence of articulate brachiopods *Leptelloidea leptelloides* (Bekk.) refers this formation to the Berezaika Horizon correlative with the upper half of the Llandeilian (the *Nemagraptus gracilis* Zone).

Ostracods are represented by single finds of *Nida nida* Sidarav., *Carinobolbina carinata estona* (Opik), *Leperditella semen* Opik, and *Ctenonotella aculeata* sp. nov.

The Kubeno Formation (lower Caradocian) is detected in depth intervals of 1570–1547 m (Well 1; core recovery of 65%) and 1568–1540 m (Well 5) in the lower part of the section. It is composed of gray to ashy gray dense marls with gray limestone beds up to 10 cm thick. The upper part of the formation mainly corresponds to dolomitized gray limestones with occasional organogenic limestone beds. The articulate brachiopods *Rafinesquina* (*Rafinesquina*) cf. *poljensis* Alich and the remains of bryozoans, crinoids, and conodonts indicate that the enclosing rocks belong to the lower half of the Caradocian.

Ostracods were found only in limestone intercalations. The following ostracods were identified in this part of the section: *Bolbina rakverensis* Sarv, *B. latimarginat* (Bonnema), *Sigmoopsis rostrata* (Krause), *S. cornuta* (Krause), and *Primitiella molli* (Bonnema). It can readily be noted that only the last two species of this undiverse assemblage were found in the underlying part of the section, while *Sigmoopsis rostrata* (Krause), *Bolbina rakverensis* Sarv, and *Bolbina latimarginata* (Bonnema) appear for the first time at this level.

The Mitino Formation (Mitino Horizon) is detected in the depth intervals of 1547–1500 m (Well 1; core recovery of 70%) and 1540–1495 m (Well 5). In its lower part, this formation is represented by alternating

Plate I

brown-gray and ash gray silts and marls that are diversely calcified; in the upper part, it mainly consists of gray to dark gray dolomites, which occasionally are brown owing to the presence of Fe-hydroxides. In the dolomites, there is a bed of gray to dark gray, dense, pelitomorphic limestone. Articulate brachiopods were not detected.

Unfortunately, owing to intensive dolomitization of the rocks, the samples recovered from these intervals contain almost no ostracod remains. Only on certain rock surfaces could ostracod casts, apparently, of *Bolbina* aff. *rakverensis* Sarv and *Conchoprimitia* sp., be found.

THE BIOSTRATIGRAPHIC ANALYSIS OF FAUNA

We made an attempt to assess the biostratigraphic implications of the studied faunal groups. The ostracod remains are irregularly distributed throughout the section. An abundant ostracod assemblage was found in the Middle Ordovician deposits, while the Lower Ordovician beds yielded a lesser number of the remains. The amount of ostracods from the Upper Ordovician section was still less.

The distribution of articulate gastropods in the studied sections is also irregular. The carbonate deposits of the Lower and Middle Ordovician contained a plentiful brachiopod assemblage, but the Upper Ordovician deposits contain almost no articulate brachiopods, thus implying a specific, probably nonmarine, sedimentation environment.

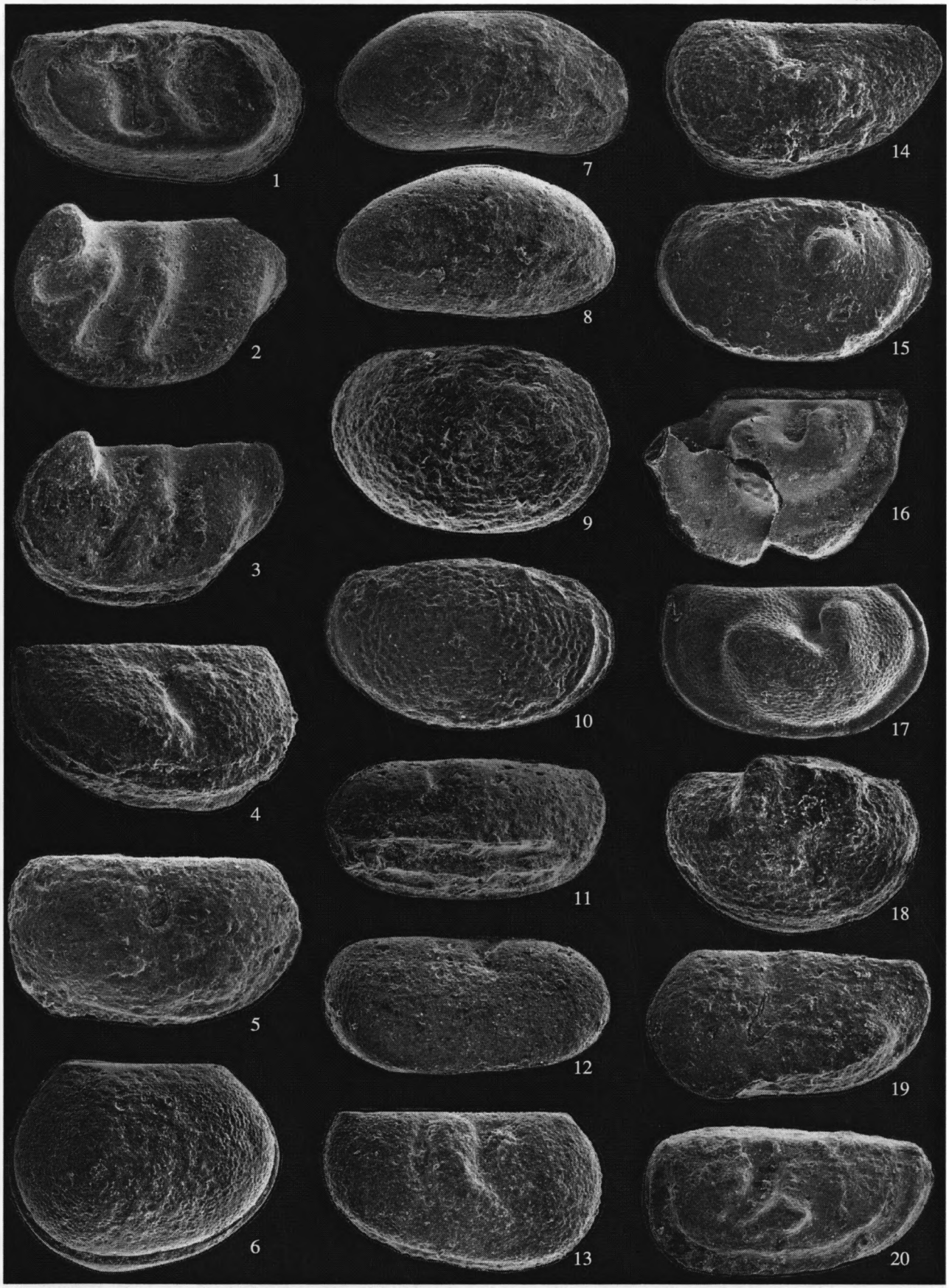
As noted above, the upper Bugino Subformation lacks ostracod remains. The Nefedovo and Veksino formations contain the same assemblage of ostracod species. Therefore, the Volkhov and Kunda horizons, reached by Gavrilov Yam wells, could not be subdivided on the basis of ostracod remains. Among the species of the Nefedovo–Veksino assemblage listed above, except the new *Piretopsis* (*Protallinnella*) *gavrilovensis*, only three forms, such as *Conchoprimitia glauconitica* (Kummerow), *C. gammae* Opik, and *Piretopsis* (*Protallinnella*) *grewingkii* (Bock), belong to the Volkhov and Kunda horizons of the Baltic region, while others were found only in the Kunda Horizon of the same region.

It is remarkable that the studied ostracod assemblage has no zonal index species, such as *Tallinnellina primaria* (Opik) of the Volkhov Horizon and *Pinnatulites procerus* (Kummerow) of the Kunda Horizon. These species were considered as zonal for the first time by Sidaravichiene (1976) and can be actually traced not only in the sections in Estonia, Lithuania, Latvia, and the Leningrad region (Sarv *et al.*, 1985), but in some of the sections of the Moscow syneclise (Prokof'ev and Kuznetsov, 1982), as indicated by Sarv. The authors, however, do not clearly specify the actual wells drilled in the Moscow syneclise, the material from which was used by Sarv in his studies. There are no data, therefore, for an analysis to elucidate the taxonomic difference between the Lower Ordovician ostracods recovered from the different wells of the Moscow syneclise. The Nefedovo–Veksino ostracod assemblage disappears at the depth of 1614 m. Higher up in the section, i.e., at the boundary between the Lower and Middle Ordovician, the ostracod composition is almost completely new.

Plate 1. The Ordovician ostracods from Gavrilov Yam Wells.

(1) *Ogmoopsis vesperi* Sarv; sample no. 4475/111, the right-hand valve, $L = 0.70$ mm; Gavrilov Yam Well 1, depth 1627.2 m; Lower Ordovician, Kunda Horizon, Veksino Formation. (2) *Protallinnella gavrilovensis* Melnikova, sp. nov.; holotype no. 4475/16, the right-hand valve, $L = 1.75$ mm; Gavrilov Yam Well 1, depth 1630.7 m; Lower Ordovician, Volkhov Horizon, Nefedovo Formation. (3) *Conchoprimitia distincta* Sarv; sample no. 4475/206, the left-hand valve, $L = 1.40$ mm; Gavrilov Yam Well 1, depth 1624.5 m; Lower Ordovician, Kunda Horizon, Veksino Formation. (4) *Conchoprimitia gammae gammae* Opik; sample no. 4475/201, the right-hand valve, $L = 1.95$ mm; Gavrilov Yam Well 1, depth 1597.9 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (5) *Conchoprimitia micropunctata* Hessland; sample no. 4475/211; (5a) the right-hand valve, $L = 1.58$ mm; (5b) an enlarged fragment of the surface of the valve with cribrate pores; Gavrilov Yam Well 5, depth 1584.5 m; Middle Ordovician, Berezaika Horizon, Skalino Formation. (6) *Tvaerenella* aff. *modesta* Sarv; sample no. 4475/46; (6a) the right-hand valve, $L = 1.05$ mm; (6b) the ventral side; Gavrilov Yam Well 1, depth 1627.5 m; Lower Ordovician, Kunda Horizon, Veksino Formation. (7) *Glossomorphites grandispinosa* (Hessland): sample no. 4475/106, side view of the valve, $L = 1.15$ mm; Gavrilov Yam Well 1, depth 1634.5 m; Lower Ordovician, Volkhov Horizon, Nefedovo Formation. (8) *Disulcina alta* Melnikova, sp. nov.; sample no. 4475/6, of the left-hand valve, $L = 1.50$ mm; Gavrilov Yam Well 5, depth 1583.5 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (9) *Ctenonotella aculeata* Melnikova, sp. nov.; holotype no. 4475/1, the left-hand valve, $L = 1.10$ mm; Gavrilov Yam Well 1, depth 1575.9 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (10) *Lunkiella suavis* Melnikova, sp. nov.; holotype no. 4475/21, the shell of a female, the left-hand valve, $L = 1.05$ mm; (10a) the dorsal side, (10b) the lateral side; Gavrilov Yam Well 1, depth 1595.9 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (11) *Comptusiella recta* Melnikova, sp. nov.; holotype no. 4475/31, the right-hand valve, $L =$ 0.7 mm; (11a) the dorsal side, (11b) the lateral side; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (12, 13) *Tallinnella discreta* Melnikova, sp. nov.; (12) holotype no. 4475/11, the right-hand valve, $L = 2.1$ mm; Gavrilov Yam Well 5, depth 1592.3 m; Middle Ordovician, Polomet' Horizon, Lunka Formation; (13) sample no. 4475/12, the ventral side, $L = 2.0$ mm; Gavrilov Yam Well 5, depth 1588.5 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (14) *Tallinnella angustata* (Krause); sample no. 4475/81, the right-hand valve, $L = 1.95$ mm, Gavrilov Yam Well 5, depth 1592.3 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (15) *Tallinnella reticulata* Sarv; sample no. 4475/86, the left-hand valve, $L = 1.85$ mm; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (16) *Tetrada* (*Perplana*) *perplana* (Neckaja); sample no. 4475/246, the left-hand valve, $L = 0.77$ mm; Gavrilov Yam Well 1, depth 1609.9 m; Middle Ordovician, Polomet' Horizon, Lunka Formation.

Plate II



The Lunka ostracod assemblage includes new members: *Ctenonotella aculeata* sp. nov., *Tallinnella discreta* sp. nov., *Disculcina alta* sp. nov., *Lukiella suavis* gen. et sp. nov., and *Comptusiella recta* gen. et sp. nov. The remaining ostracod species, typical of the Lunka assemblage and recovered from Gavrilov Yam Wells, were recovered for the first time in the territory of the Baltic region, except *Levisulculus lineatus* Jaanusson, which is known only from the Ordovician deposits of Sweden. The most numerous in this assemblage are *Nida nida* Sidarav., *Sigmoopsis platyceras* (Opik), *Tetrada (Preplana) preplana* (Neckaja), *Tallinnella angustata* (Krause), *Tallinnella discreta* sp. nov., *Leperditella semen* Opik, *Longiscula arcuaria* Neckaja, and *Lunkiella suavis* sp. nov.

In the Skalino Formation, the ostracods are scarce, because the larger part of rocks in this interval is dolomitized.

It is of interest to note the appearance of *Bolbina rakverensis* Sarv in the Gavrilov Yam section beginning from the lower Caradocian level (Kubeno Formation). Previously, Sarv (Põlma *et al.*, 1977) intimated that this species is typical of the Oandu–Rakvere time in the late Caradocian.

There are almost no ostracods in the deposits of the Mitino Formation.

An analysis of the list of the ostracod species discussed above shows that, apart from the new species, they were known from the Baltic region, Norway, and Sweden and from the erratic boulders of Central Europe. A close connection between the respective basins is indicated by the existence in a large number of the Middle Ordovician ostracod species common for the Moscow syneclise and Baltoscandia.

Of particular interest is the appearance of certain species at an earlier level in the Gavrilov Yam section. In addition to the *Bolbina rakverensis* Sarv species discussed above, these are *Nida nida* Sidarav., *Sigmoopsis cornuta* (Krause), *S. rostrata* (Krause), *Leperditella semen* Opik, *Pseudostrepula acuta* (Bonnema), *Primitiella molli* (Bonnema), *Pentagona joehviensis* (Sarv), and *Sigmobolbina variolaris* (Bonnema). Perhaps, this is because these taxa first appeared in the basin that existed on the territory of the Moscow syneclise, and, at a later period, they migrated to the west.

DESCRIPTION OF OSTRACODS

Family Ctenonotellidae E. Schmidt, 1941

Subfamily Ctenonotellinae E. Schmidt, 1941

Genus *Ctenonotella* Opik, 1937

Species *Ctenonotella aculeata* Melnikova, sp. nov.
Plate 1, no. 9

The name of the species is derived from the Latin *aculeata* meaning spiny.

Holotype: PIN, no. 4475/1, shell; Moscow syneclise, Gavrilov Yam Well 1, depth 1575.9 m; Middle Ordovician, Berezaika Horizon, Skalino Formation.

Material. About ten valves and three shells satisfactorily and well preserved.

Plate 2. The Ordovician ostracods from Gavrilov Yam Wells.

(1) *Nida nida* Sidaravichene; sample no. 4475/251, the right-hand valve, $L = 1.15$ mm; Gavrilov Yam Well 5, depth 1605.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (2) *Sigmoopsis rostrata* (Krause); sample no. 4475/131, the left-hand valve, $L = 1.1$ mm; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (3) *Sigmoopsis platyceras* (Opik); sample no. 4475/136, the left-hand valve, $L = 1.15$ mm; Gavrilov Yam Well 5, depth 1605.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (4) *Sigmobolbina variolaris* (Bonnema); sample no. 4475/161, the right-hand valve, $L = 0.85$; Gavrilov Yam Well 1, depth 1597.7 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (5) *Euprimites* cf. *eutropis* (Opik); sample no. 4475/40, the left-hand valve, $L = 1.2$ mm; Gavrilov Yam Well 5, depth 1605.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (6) *Leperditella semen* Opik; sample no. 4475/196, the right-hand valve, $L = 1.47$ mm; Gavrilov Yam Well 1, depth 1609.9 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (7) *Lonfiscula parrectis* Neckaja; sample no. 4475/231, the right-hand valve, $L = 1.1$ mm; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (8) *Longisculla arcuaria* Neckaja; sample no. 4475/236, the right-hand valve, $L = 0.95$ mm; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (9) *Baltonotella* sp.; sample no. 4475/256, the left-hand valve, $L = 1.02$ mm; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (10) *Ahlintella* ex gr. *linta* Sidaravichiene; sample no. 4475/216, the left-hand valve, $L = 0.60$ mm; Gavrilov Yam Well 5, depth 1602.8 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (11) *Severobolbina severa* (Sarv); sample no. 4475/151, side view of the left-hand valve, $L = 1.55$ mm; Gavrilov Yam Well 5, depth 1583.2 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (12) *Primitiella molli* (Bonnema); sample no. 4474/246, the right-hand valve, $L = 0.83$ mm; Gavrilov Yam Well 5, depth 1606.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (13) *Carinobolbina carina estona* Opik; sample no. 4475/156, the right-hand valve of tecnomorph, $L = 0.85$ mm; Gavrilov Yam Well 5, depth 1588.5 m; Middle Ordovician, Berezaika Horizon, Skalino Formation. (14) *Hesslandella unica* Sarv; sample no. 4475/121, the left-hand valve, $L = 0.70$ mm; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (15) *Gryphiswaldensia gryphiswaldensia* Schallreuter, sample no. 4475/186, the right-hand valve, $L = 0.65$ mm; Gavrilov Yam Well 5, depth 1585.5 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (16) *Bolbina latimarginata* (Bonnema); sample no. 4475/66, the right-hand valve, domicilium $L = 0.91$ mm; Gavrilov Yam Well 1, depth 1565.4 m; Middle Ordovician, Kubeno Formation. (17) *Bolbina rakverensis* Sarv; sample no. 4475/61, the right-hand valve, $L = 0.80$ mm; Gavrilov Yam Well 1, depth 1565.7 m; Middle Ordovician, Kubeno Formation. (18) *Klimphores* ex gr. *minimus* (Sarv); sample no. 4475/226, the right-hand valve, $L = 0.47$ mm; Gavrilov Yam Well 5, depth 1602.8 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (19) *Pentagona joehviensis* (Sarv); sample no. 4475/171, the right-hand valve, $L = 0.90$ mm; Gavrilov Yam Well 5, depth 1602.8 m; Middle Ordovician, Polomet' Horizon, Lunka Formation. (20) *Steusloffia acuta* (Krause); sample no. 4475/91, the right-hand valve, $L = 0.90$ mm; Gavrilov Yam Well 5, depth 1602.0 m; Middle Ordovician, Polomet' Horizon, Lunka Formation.

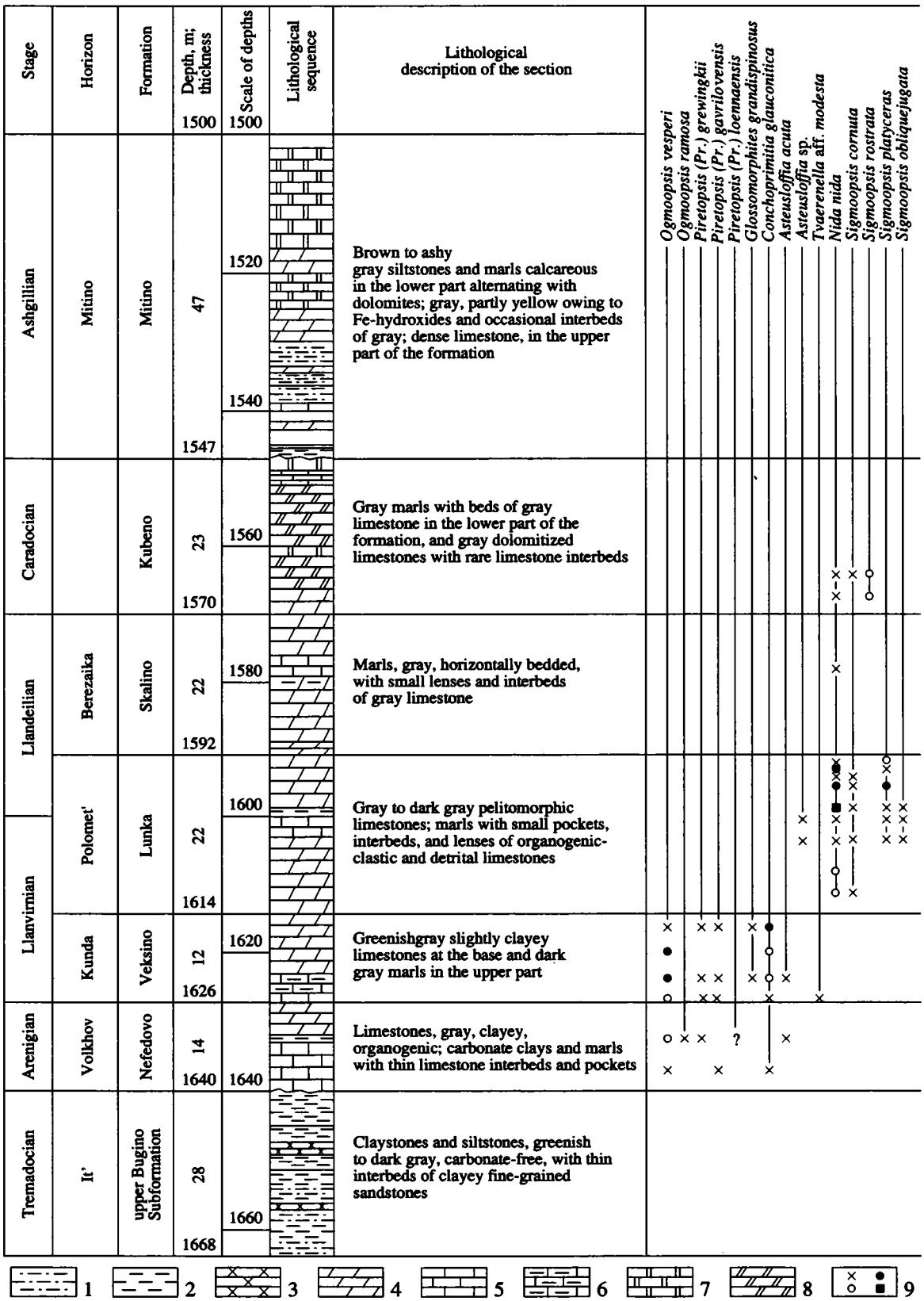


Fig. 2. Distribution of ostracod and articulate brachiopod species in the section of Gavrilov Yam Well 1 (black triangle on the inset map shows the location of Gavrilov Yam wells drilled in the Yaroslavl Region, Russia).

(1) siltstone; (2) claystone; (3) sandstone; (4) marls; (5) limestone; (6) clayey limestone; (7) dolomite; (8) dolomitic limestone; (9) number of ostracod samples: (×) 1–5; (○) 6–10; (●) 11–20; (■) more than 21.

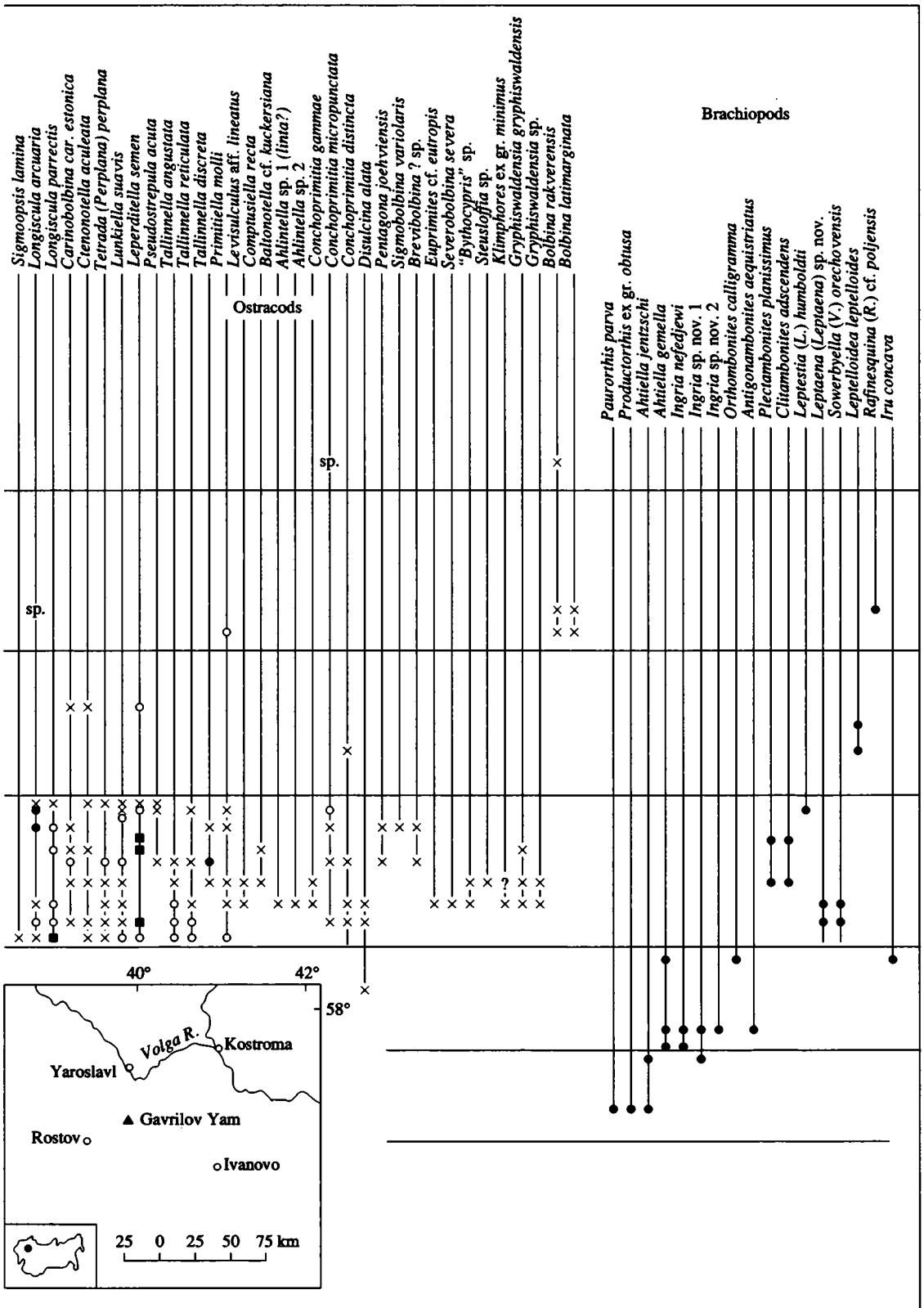

Ostracods
Brachiopods
Sigmoopsis lamina
Longiscula arcuaria
Longiscula parrectis
Carinobolbina car. estonica
Ctenonotella aculeata
Tetrada (Perplana) perplana
Lunkiella suavis
Leperditella semen
Pseudostrepula acuta
Tallinnella angustata
Tallinnella reticulata
Tallinnella discreta
Primitiella molli
Levisulculus aff. lineatus
Comptusiella recta
Baltonotella cf. kuckersiana
Ahlintella sp. 1 (linta?)
Ahlintella sp. 2
Conchoprimitia gammae
Conchoprimitia micropunctata
Conchoprimitia distincta
Disulcina alata
Pentagona joehviensis
Sigmobolbina variolaris
Brevibolbina ? sp.
Euprimites cf. eutropis
Severobolbina severa
"Bythocypris" sp.
Steusloffia sp.
Klimphores ex gr. minimus
Gryphiswaldensia gryphiswaldensis
Gryphiswaldensia sp.
Bolbina rakverensis
Bolbina latimarginata
sp.
sp.
Paurorthis parva
Productorthis ex gr. obtusa
Ahtiella jentzschi
Ahtiella gemella
Ingria nefedjewi
Ingria sp. nov. 1
Ingria sp. nov. 2
Orthombonites calligramma
Antigonambonites aequistriatus
Plectambonites planissimus
Clitambonites adscendens
Leptestia (L.) humboldti
Leptaena (Lepitaena) sp. nov.
Sowerbyella (V.) orechovensis
Leptelloidea lepiellоides
Rafinesquina (R.) cf. poljensis
Iru concava
40°
42°
58°
Volga R.
Kostroma
Yaroslavl
Gavrilov Yam
Rostov
Ivanovo
25 0 25 50 75 km

Fig. 2. (Contd.)

Diagnosis. The bilobate shell is of subpreplete outline. The forelobe L2 is almost straight; L3 is longer and wider than L2, and it slightly curves in the middle part. The velar structure has the shape of a thin cicatrixlike ridge. The surface of the shell is finely tubercular.

Description. The shell is large, subpreplete in shape, moderately convex, and bilobate. The anterior end is high and well rounded; the posterior end slopes in the posteroventral part. The anterodorsal angle is obtuse; the posterodorsal angle narrows down to a spine. Both lobes are developed in the anterior half of the valves. The anterior lobe L2 is moderately convex, almost straight. L3 is longer and wider than L2; it slightly bends in the middle part. The sulcus between the lobes (S2) is wide and shallow. Behind L3, a deeper sulcus stretches along its full length; it is narrower and longer than S2. There are minute spines along the anterior end. The maximal height and convexity are in the anterior one third of the shell. The surface of the shell is finely tubercular. The velar structure is represented by a very thin cicatrixlike ridge stretching from the anterodorsal angle to the middle of the posterodorsal margin.

Dimensions, mm	l	h	w
Holotype no. 4475/1	1.1	0.7	0.6
Specimen no. 4475/2	0.6	0.35	–

Variability is manifested in the degree of tapering of the posterior end and in the slight changes in the height of the lobes.

Comparison, remarks. The species differs from the nearest *Ctenonotella elongata* Opik. 1937, recovered from the Kukruse Horizon (CII) in Estonia, by the tapering spiniform posterior end and from *C. bidens* (Krause, 1892), with a similar structure of the lobes, by a different velar structure. The velar ridge of *C. bidens* is rather wide, with traces of radiancy, whereas the new species has it in the form of a small cicatrix. However, the cicatrixlike aspect of the velar ridge is, probably, a result of poor preservation.

Geological age and geographic range. Middle Ordovician; Polomet' and Berezaika horizons; Moscow syneclise.

Family Tetradellidae Swartz, 1936

Subfamily Sylthinae Schallreuter, 1982

Genus *Disulcina* Sarv, 1959

Species *Disulcina alta* Melnikova, sp. nov.

Plate 1, no. 8

The name of the species is derived from the Latin *alta* meaning broad.

Holotype: PIN no. 4475/6, the left valve; Moscow syneclise, Gavrilov Yam Well 5, depth 1604.5 m; Middle Ordovician, Polomet' Horizon, Lunka Formation.

Material. Seven well-preserved valves.

Diagnosis. The shell is of an amplete shape; it is high, with two sulci (S2 and S3). S2 is straight, vertical, and open toward the dorsal margin. The shape of S3 is irregular droplike. The surface of the shell is porous, and that of the inner gently inclined slopes of the sulci is striated.

Description. The shell is middle-sized, amplete in shape, moderately convex, and high. The dorsal margin is straight and long; the ventral margin is plano-convex and arcuate. The frontal end is somewhat higher than the posterior end, which slightly slopes in the posteroventral part.

On the lateral surface of the valve, there are two wide and moderately deep sulci. The frontal sulcus (S2) is straight, vertical, and open towards the dorsal margin; it is situated somewhat nearer to the frontal end and begins below the dorsal margin and nearly reaches the middle height of the shell. The second sulcus (S3) is close to S2 and somewhat longer. It has an irregular droplike shape; this sulcus is very narrow in the dorsal part and widens toward the posterior end. Along the posterior end of the shell, there is a narrow lateral projection. The maximal height and convexity are located in the middle. The surface of the shell is porous, while that of the gently inclined internal slopes of the sulci is striated.

Variability. The width of S3 slightly varies.

Dimensions, mm	l	h
Holotype no. 4475/6	1.50	1.05
Specimen no. 4475/7	1.70	1.05

Comparison. The new species differs from all the known disulcinas by the very wide and closely placed sulci and by the porous surface of the shell.

Geological age and geographic range. Middle Ordovician, Polomet' Horizon; Moscow syneclise.

Subfamily Tallinnellinae Schallreuter, 1967

Genus *Tallinnella* Jaanusson, 1957

Species *Tallinnella discreta* Melnikova, sp. nov.

Plate 1, nos. 12, 13

The name of the species is derived from the Latin *discretus* meaning divided.

Holotype: PIN, no. 4475/11, the right valve; Moscow synecline, Gavrilov Yam Well 5, depth 1592.3 m; Middle Ordovician, Polomet' Horizon, Lunka Formation.

Material. About 80 well-preserved valves and shells.

Diagnosis. The shell is of a preplete shape and has four lobes. L1 and L3 are long and slightly rise above the dorsal margin. L2 is short and narrow. L4 is narrow; towards the middle part it sharply declines to complete reduction. The sulci are distinct. The velar structure has

the shape of a wide ridge. The surface of the shell is smooth.

Description. The shell is large and has a preplete shape and four lobes. The dorsal margin is straight and long; the ventral margin is arcuately bent. L1 and L3 are long, relatively wide, and high; they are thicker in the dorsal part and slightly protrude over the dorsal margin. L2 is short and narrow. L4 is lower than L1 and L3 and narrow in its ventral extremity; towards the middle part, it sharply declines to complete reduction, but towards the dorsal margin it again widens and thickens. The lower ends of the lobes merge with the junction ridge, which is relatively wide in the anterior part of the shell and narrows towards the posterior end. The sulci are distinct; S1 and S2 are small, whereas S3 is rather wide and somewhat deeper than S2. The velar structure is represented by a wide ridge, which begins at the anterodorsal angle and reaches the middle height of the posterior end. The velar edge of the shells of females is somewhat wider in the anteroventral part. Between the junction and velar ridges, there is a wide furrow. The surface of the shell is smooth.

Dimensions, mm	l	h
Holotype no. 4475/11	2.1	1.3
Specimen no. 4475/12	2.0	1.2

Variability. The width of the lobes and sulci varies.

Comparison. The species differs from all the previously known tallinnellas by the fragmented structure of L4 and from the nearest species *T. angustata* (Krause, 1891) by the smooth surface of the entire shell.

Geological age and geographic range. Middle Ordovician, Polomet' Horizon; Moscow syneclise.

Genus *Piretopsis* Henningsmoen, 1953

Subgenus *Piretopsis* (*Protallinnella*) Jaanusson, 1957

Piretopsis (*Protallinnella*) *gavrilovensis* Melnikova, sp. nov.

Plate 1, no. 2

The species *gavrilovensis* is named after the Gavrilov Yam site.

Holotype: no. 4475/16, the shell of a heteromorph; Moscow syneclise, Gavrilov Yam Well 1, depth 1630.7 m; Lower Ordovician, Volkhov Horizon, Nefedovo Formation.

Material. About 20 well-preserved valves and shells.

Diagnosis. The shell is amplete in shape; L1, L3, and L4 reach the dorsal margin. L1 is ridgelike; L2 is short; L3 is wide and thicker in the dorsal part; L4 is very narrow and ridgelike. The valves are joined by a narrow junction ridge. The velar ridge has the shape of a narrow compact ridge with fine striation along the outer margin. The surface of the shell is ornamented with middle-sized nodes.

Description. The shell is large, amplete in shape, moderately convex, and tetralobate. L1, L3, and L4 reach the dorsal margin. L1 is ridgelike in the lower part and widens in the dorsal part. L2 is narrow, short, and begins considerably lower than the dorsal margin. L3 is wide, thicker in the dorsal part. L4 is very narrow and ridgelike. The dorsal extremity of L4 bends and stretches along the dorsal margin to the posterodorsal angle; it slightly subsides along the posterior margin of the valve. The lobes are joined at the base by a narrow junction ridge, which turns upwards at the base of L3. S2 is the deepest, and S3 is the widest and longest. The velar structure is represented by a narrow compact ridge with a fine striation along the outer margin. The velar ridge stretches from the anterodorsal angle to the lower end of the posterior margin. On the shells of heteromorphs, the velar ridge in the anteroventral part is slightly distended. There is a fine cicatrix between the velar ridge and the margin of the valves. Minute spines are occasionally present along the marginal rim of the posterior margin of the valves. Except the smooth and slightly porous lobes, the shell is ornamented with middle-sized nodes.

Dimensions, mm	l	h	w
Holotype no. 4475/16	1.75	1.15	–
Specimen no. 4475/17	1.85	1.15	–
Specimen no. 4475/18	1.70	1.10	0.5

Variability. The width of the lobes may vary.

Comparison. The new species resembles the genoholotype *Piretopsis (Pr.) grewingkii* (Bock, 1867) by the structure of L4 but differs from it by thicker L1 and L3 and by the narrower velar ridge. It differs from the near species *Piretopsis* (Pr.) *loennaensis* Sarv, 1963, by the distended dorsal parts of L1 and L3 and by the curve of the junction ridge upwards only at the base of L3.

Geological age and geographic range. Lower Ordovician, Volkhov and Kunda horizons; Moscow syneclise.

Family ? Tetradellidae Swartz, 1936

Genus *Lunkiella* Melnikova gen. nov.

The genus *Lunkiella* is named after the Lunka River in the Yaroslavl Region.

Genotype *Lunkiella suavis* sp. nov.; Middle Ordovician, Polomet' Horizon, Lunka Formation; Moscow syneclise, Gavrilov Yam Well.

Diagnosis. The shell is preplete in shape, moderately convex, trilobate, and inequivalve. The frontal lobe is formed by the merged L1 and L2. The anterior and posterior lobes fringe the ends of valves. L3 is almost vertical in its upper part and bends in the middle toward the frontal edge. The dorsal extremities of lobes are extended and thickened; they look subsquare from the side of the dorsal margin. The lobes in the ventral part are conjoined through the junction ridge. The sulci

are wide, flat, and open towards the dorsal margin. The velar structure of the female shells is represented by a narrow ridge stretching along the frontal edge without separation from the domicilium surface. The subvelar field is subconcave and divided in the upper part by a septum forming a rounded loculus near the anterodorsal angle on each valve. The velar ridge of the tecnomorph shells is practically absent, and the loculus is not formed.

Comparison, remarks. By the type of dissection of the lateral surface, the new genus is close to the subgenus *Tetrada* (*Perplana*) Sidaravichiene, 1992 of the Tetradellidae Family. The tetradellids, however, have the tendency to form loculi in the supervelar antrum. The loculi on the shells of the new genus are in the subvelar field. Therefore, the reference of *Lunkiella* genus to the Tetradellidae Family is tentative.

Specific composition. Type species.

Geological age and geographical range. Middle Ordovician, Moscow syneclise.

Lunkiella suavis Melnikova, sp. nov.
Plate 1, no. 10

The name of the species is derived from the Latin *suavis* meaning pleasant, attractive.

Holotype: PIN, no. 4475/21, a heteromorph shell; Moscow syneclise, Gavrilov Yam Well 1, depth 1597.9; Middle Ordovichian, Polomet' Horizon, Lunka Formation.

Material. About 50 well-preserved valves and shells.

Diagnosis. The shell is subpreplete in shape, trilobate, and inequivalve. The lobes are high and ridgelike, with different width. The dorsal extremities of all valves are thickened and subsquare in shape on the dorsal side. The velar structure on the heteromorph shells is represented by a narrow velar ridge. The subvelar field is subconcave and divided by a septum forming a rounded loculus. The surface of the shell is smooth.

Description. The shell is of middle size, subpreplete in shape, moderately convex, trilobate, and inequivalve; the left valve is somewhat larger than the right one and envelops the latter along the ventral margin. The dorsal margin is long, orthocuneate. The dorsal angles are obtuse; the anterior angle is larger than the posterior one. The lobes are high and ridgelike with different width. The frontal lobe L1 is wide and merges with L2, which is scarcely noticeable in just a few certain specimens. L1 and L4 frame the ends of valves. L3 is located in the middle; in the dorsal part, it has a triangular distention; nearer to the middle, it narrows and bends towards the anterior end. L4 has a lanceolate distention in the dorsal part. The dorsal extremities of all the lobes are thickened and look subsquare from the dorsal side. The lobes in the ventral part are conjoined through the junction ridge of a moderate width. The sulci between the lobes are wide and shallow; they have a flattened bottom and open toward the dorsal margin. The surface of the shell is smooth. The velar structure on the heteromorph shell is represented by a velar ridge along the anterior end, which is not separated from the surface of the domicilium. The subvelar field is subconcave; it is separated by a septum and forms a rounded loculus near the anterodorsal angle. There is no ridge on the shells of heteromorphs.

Dimensions, mm	l	h	w
Holotype no. 4475/21 ♀	1.05	0.65	0.30
Specimen no. 4475/22 ♂	0.95	0.60	0.36
Specimen no. 4475/23 ♀	1.1	0.65	0.30
Specimen no. 4475/24 ♀	1.1	0.68	0.28

Variability. The width of lobe L4 changes from uniformly wide along its entire length to narrowing at the ventral extremity. Moreover, there are slight changes in the thickening of the dorsal extremities of the lobes.

Geological age and geographic range. Middle Ordovician, Polomet' Horizon; Moscow syneclise.

Family ? Monotiopleuridae Guber et Jaanusson, 1965

Genus *Comptusiella* Melnikova, gen nov.

The species name is derived from the Latin *comptus* meaning beautiful, elegant.

Genotype: *Comptusiella recta* gen. et sp. nov., Middle Ordovician, Polomet' Horizon, Lunka Formation of the Moscow syneclise, Gavrilov Yam Well.

Diagnosis. The shell is elongated-rectangular in shape, low, plano-convex, and inequivalve; the left valve envelops the right one along the entire free margin. The dorsal margin is long and located in the hollow formed by the raised and thickened dorsal margins of the valves. There is a thin ridge on the lateral surface of the shell near the ventral margin. The surface of the shell is smooth.

Comparison and remarks. The elongated-rectangular outline of the shell of the representatives of the new species indicates a certain likeness with such genera as *Primtiella* Ulrich and *Foveaprimitiella* Schallreuter, 1972. However, such a feature of the new genus as the position of the dorsal margin in a recess clearly distinguishes it from all the known monotiopleurids. The reference of the new species to the Monotiopleuridae Family is, therefore, conventional.

Specific composition. Type species.

Geological age and geographic range. Middle Ordovician, Moscow syneclise.

Comptusiella recta Melnikova, sp. nov.
Plate 1, no. 11

The species name is derived from the Latin *recta* meaning straight.

Holotype: PIN, no. 4475/31, a shell; Moscow syneclise. Gavrilov Yam Well 5, depth 1602 m; Middle Ordovician, Polomet' Horizon, Lunka Formation.

Material. Five well-preserved shells.

Diagnosis. The shell is elongated-rectangular in shape, plano-convex, and inequivalve. The dorsal margin is straight, and its length is somewhat less than that of the shell; it is located in a narrow recess. The ventral margin is almost parallel to the dorsal margin. On the side surface, there is a thin ridge parallel to the ventral margin. The surface of the shell is smooth.

Description. The shell is small, elongated-rectangular, low, plano-convex, and inequivalve; its left valve envelops the right one along the free margin. The covering is irregular, considerable along the ventral margin and less significant along the anterior and posterior ends. The dorsal margin is straight and long; it is slightly shorter than the shell. The dorsal margin is located in a narrow recess formed by the elevated and slightly thickened dorsal areas of the valves. The ventral margin is subconvex and almost parallel to the dorsal margin. The anterior and posterior ends have equal height and similar contours. There is a thin ridge on the lateral surface of the valves, which runs, with a slight retreat, parallel to the ventral margin. This ridge stretches from the frontal to the posterior end. The surface of the shell is smooth.

Dimensions, mm	l	h	w
Holotype no. 4475/31	0.7	0.3	0.2
Specimen no. 4475/32	0.7	0.3	0.2

Variability. The species display permanent features.

Geological age and geographic range. Middle Ordovician, Polomet' Horizon; Moscow syneclise.

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Reviewer A.S. Alekseev

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The Upper Carboniferous Foraminiferal Zonation in the North Uralian Region

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Abstract—New data on the vertical and lateral distribution of key fusulinid species revealed peculiar features of their communities in the North Uralian region. The Kasimovian Stage includes here all zonal species known in the Urals and East European platform, whereas regional fusulinids are characteristic of the Gzhelian Stage. The upper stage of the Carboniferous incorporates only the *Triticites rossicus* and *Daixina tschernovi* zones. The Carboniferous–Permian boundary is placed at the base of the *Daixina robusta* Subzone attributed to the lower part of the standard *Schwagerina fusiformis–Sch. vulgaris* Zone.

Key words: *zones, index species, Kasimovian Stage, Gzhelian Stage, western slope of the Urals, Chernyshev Ridge, Pechora syneclise.*

In this work, we consider the North Uralian folded region as comprising a part of the Urals extending northward of the Un'ya River valley and the Chernyshev Ridge representing the transverse structure of the Uralian foredeep. A series of reference sections was studied in these and adjacent structures of the Pechora syneclise (Fig. 1).

The first data on Carboniferous foraminifers from the Urals were published in the late 1930s by Rauser-Chernousova *et al.* (1936) and Varsanof'eva (1940), who studied the collections of A.A. Chernov, T.A. Dobrolyubova, and E.D. Soshkina obtained when they mapped sheets 122 and 123 of the Geological Map of Russia.

In the 1950s and afterward, foraminifers became the leading fossils used in stratigraphy of Carboniferous sections. The Upper Carboniferous deposits of our region were studied during the large-scale geological survey by L.N. Belyakov, V.S. Enokyan, V.P. Lipatov, G.R. Proskurin, B.I. Tarbaev, and other geologists of various institutions.

Special attention is devoted to the Upper Carboniferous stratigraphy in many publications (Eliseev, 1958, 1960,1963; Eliseev and Mikhailova, 1962; Mikhailova, 1971, 1974; Chermnykh, 1960, 1976; Chermnykh *et al.*, 1970, 1974; Murav'ev, 1968; Kalashnikov and Mikhailova, 1971, 1984). In this period, Mikhailova (1974) performed the classical study on fusulinids and suggested the succession of *Protriticites pseudomontiparus–Obsoletes obsoletus*, *Montiparus montiparus*, and *Triticites quasiarcticus–T. acutus* zones for the Kasimovian Stage. In the Gzhelian Stage, she distinguished the *Triticites rossicus* Beds and *Daixina sokensis* Zone.

These data proved that zonal index species of the Kasimovian Stage stratotype that are accepted in uniform stratigraphic schemes for the Urals and Russian platform also occur in the North Uralian region. At the same time, the Gzhelian Stage was shown to be peculiar with respect to the vertical distribution of fusulinids, which suggested that the *Triticites rossicus* Beds of North Uralian sections correspond to the standard *Triticites stuckenbergi* and *Jigulites jigulensis* zones.

During the subsequent study, the reference Upper Carboniferous sections were reexamined two or three times with the goal of collecting fusulinids. New finds and additional revision of available materials and thin-sections allowed us to refine the distribution patterns of index genera and species and to establish the more correct time succession of fusulinid assemblages in respective beds.

In this work, we consider the term "zone" as corresponding to the stratigraphic body (three-dimensional in space) delineating the occurrence range of an index taxon and restricted on the top by the zone of the younger index taxon. As an exception, the zone can be distinguished even in a section lacking zonal taxa, provided that it bears associated forms and is located in the distribution area of the index species.

RESULTS

Exposure 80 at the Shar'yu River (the southern part of the Chernyshev Ridge) is one of the reference sections of the Upper Carboniferous in the North Uralian region. According to Eliseev, this section exhibits (from the base upward) the *Fusulinella usvae* Beds successively overlain by the *Triticites montiparus*, *Triticites arcticus–T. acutus*, and *Triticites rossicus* beds. Above

them, there is the Asselian Stage, whose lower unit corresponds to the *Pseudofusulina anderssoni–Daixina* Beds correlated by Eliseev with the *Daixina sokensis* Zone, the horizon with pseudofusulinids in his scheme of 1951 (Eliseev, 1963).

Zonal taxa of the Kasimovian Stage. In the Shar'yu reference section, Mikhailova (1974) distinguished, as noted at the beginning, all three fusulinid zones of the Kasimovian Stage accepted in the standard stratigraphic schemes for the Urals (*Reshenie ...*, 1951) and Russian platform (*Reshenie ...*, 1965).

Reexamination of geological collections and new thin-sections for the Shar'yu locality showed that Sample 839b (Fig. 2) contains *Usvaella usvae* (Dutk.), whereas *Obsoletes biconicus* Kir., *Protriticites pseudomontiparus* Put., *Pr. plicatus* Kir., *Pr. globulus* Put., *Pr. ovatus* Put., and *Fusulinella bocki* Moell. are present in Sample 151.

In addition to index species (identified in Samples 153, 379, and 844c), the overlying *Montiparus montiparus* Zone yielded other representatives of this genus (Fig. 2): *Montiparus umbonoplicatus* (Raus. et Bel.) from Samples 153, 379, 844c, and 845c; *M. subcrassulus* (Ros.) from Samples 153, 844c, and 845c; *M. tenebrosus* (Z. Mikh.) from Samples 153 and 844c; *M. sinuosus* (Ros.) from Samples 153, 844c, and 845c; *M.* cf. *paramontiparus* (Ros.) from Samples 153 and 844c; and *M. umbonoplicatus* (Raus. et Bel.).

The upper *Triticites quasiarcticus–T. acutus* Zone of the Kasimovian Stage (sampling points 154–158) was also distinguished in the reference section on the basis of its index species. Samples 154, 381, and 841d representing the common sampling level yielded *Triticites quasiarcticus* Sol., *T. acutus* Dunb. et Cond., *T. petschoricus* Raus. et Bel., *T. petschoricus brevis* Raus. et Bel., *T. petschoricus varsanofievae* Z. Mikh., *T. subacutus* Z. Mikh., *T. simplex* Schellw., *T. secalicus* Say, *T. noinskyi* Raus. et Bel., *T.* cf. *fortissimus* Raus., *T. variabilis* Ros., *Montiparus subcrassulus* (Ros.), *M. sinuosus* (Ros.), *M. sinuosus boreus* Z. Mikh., and *Usvaella usvae* (Dutk.). The lower boundary of this zone was placed at the top of the limestone bed bearing solitary corals and overlain by the fusulinid limestone containing a diverse assemblage of fossils, whose *Triticites* species are listed above. The zonal index species are present in Samples 156, 157, and 22 (Fig. 2).

Much further to the south of the reference section, at the western slope of the northern Urals (the Podcherem River, Morchanov-Kyrta area, Fig. 3), there is another section of the Kasimovian Stage. The Middle–Upper Carboniferous boundary is unexposed here, and the lowermost unit is represented by detrital fusulinid limestones of the *Protriticites pseudomontiparus–Obsoletes obsoletus* Zone.

Samples 3–13 and 438–440 from this section yielded the following index species: *Protriticites subschwagerinoides* Ros., *P. pseudomontiparus* Ros., *Obsoletes* cf. *biconicus* Kir., *Dagmarella ylychensis* (Raus.), *Eowaeringella* sp., and *Pulchrella pulchra* (Raus. et Bel.).

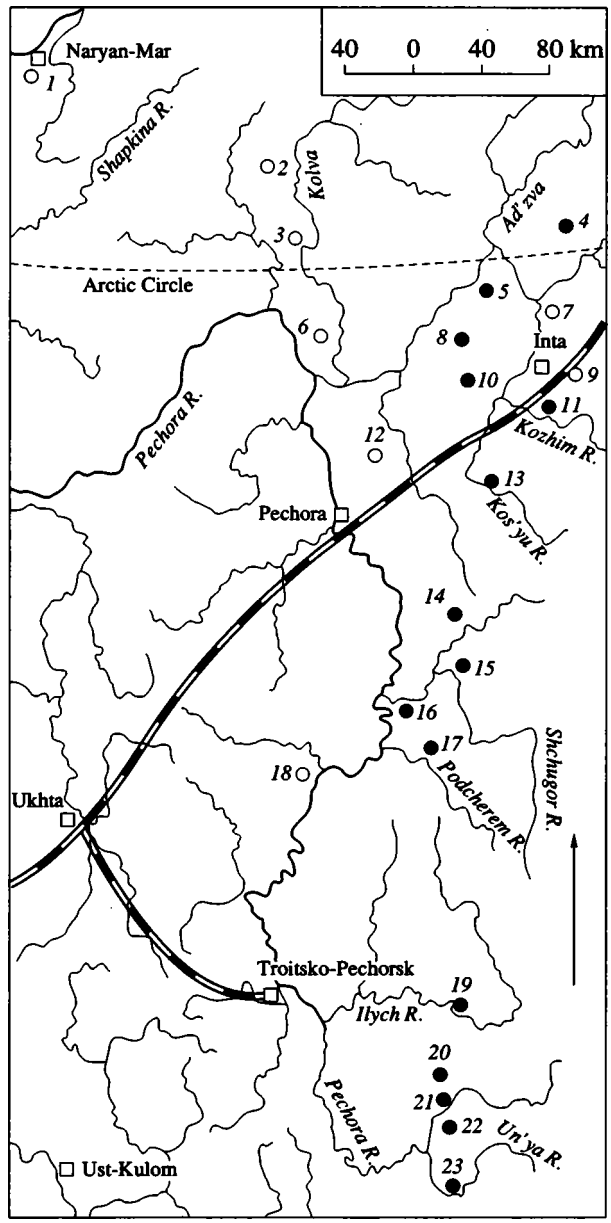
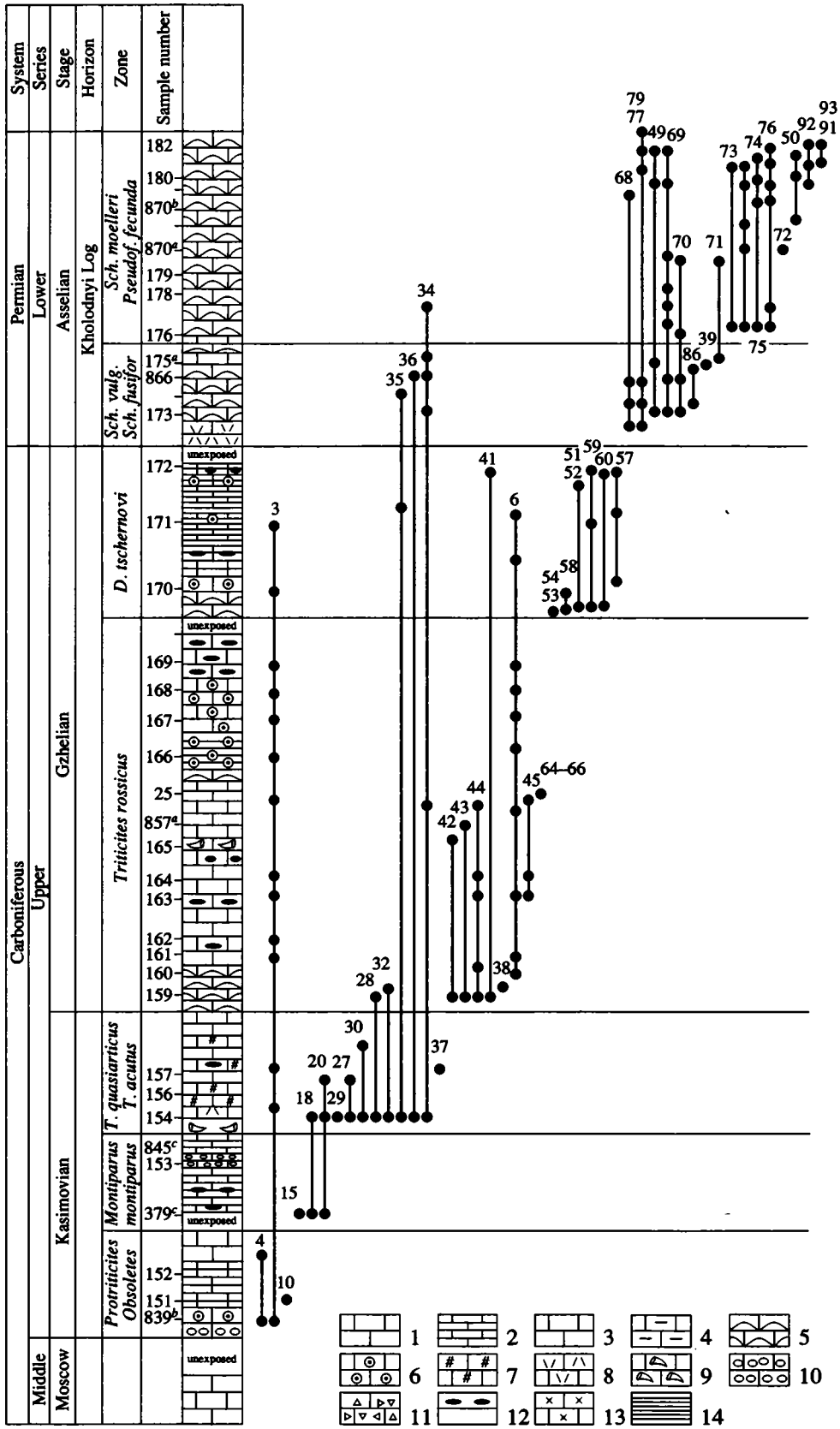


Fig. 1. Geographic localities of wells (open circles) and exposures (solid circles): (*1*) Naryan-Mar-1; (*2*) Khar'yaga-1, -47, and -61; (*3*) Vozei-102; (*4*) Ul'vozh; (*5*) Zaostrennaya-6, -7, and -8; (*6*) Usa–Kolva-4, -6, and -13; (*7*) Kochmes-1; (*8*) Shar'ya-80; (*9*) Inta; (*10*) Bol'shaya Sar'yuga; (*11*) Kozhim; (*12*) Synya-2 and -4; (*13*) Kos'yu; (*14*) Gerd'yu-4; (*15*) Verkhnie Vorota; (*16*) Nizhnie Vorota; (*17*) Podcherem-47 and -49; (*18*) Srednii Vuktyl-23; (*19*) Ilych (Moleben-El'); (*20*) Bol'shaya Shaitanovka; (*21*) Verkhnyaya Pechora; (*22*) Utlan; (*23*) Un'ya-28.

Data suggesting that the last locality also includes the younger *Montiparus montiparus* Zone are unavailable at present.


System
Series
Stage
Horizon
Zone
Sample number
Permian
Carboniferous
Lower
Upper
Middle
Asselian
Gzhelian
Kasimovian
Moscow
Kholodnyi Log
Sch. moelleri Pseudof. fecunda
Sch. vulg. Sch. fusifor
D. tschernovi
Triticites rossicus
T. quasiarticus T. acutus
Montiparus montiparus
Protriticites Obsoletes
unexposed
1 2 3 4 5
6 7 8 9 10
11 12 13 14

The higher *Triticites quasiarcticus–T. acutus* Zone of the locality is represented by detrital fusulinid limestones bearing its index species associated with *Triticites subacutus* Z. Mikh., *T. schwageriniformis* Raus., *T. noinskyi plicatus* Ros., *T. petschoricus* Raus. et Bel., *T. petschoricus brevis* Raus. et Bel., *T. petschoricus varsanofievae* Z. Mikh., and other fusulinids (Samples 15–18 and 446–451).

Unfortunately, other sections of the Chernyshev Range and western Urals are usually stratigraphically incomplete, except for some localities at the Zaostrennaya, Shchugor, and Verkhnyaya Pechora rivers (Fig. 1), where the Kasimovian deposits seem to be fairly representative. In the western unexposed areas, these deposits were locally recovered by deep drilling (e.g., the Synya locality, wells 2 and 4); however, the core samples obtained are fragmentary and do not characterize the Kasimovian Stage in its full scope (Konovalova, 1991).

In general, our analysis demonstrated that distribution patterns of indicative Late Carboniferous fusulinid genera and species in the studied region are irregular. For instance, the Podcherem-49 section (the Oselka-Kyrta area) is continuous and rich in fusulinids of the *Protriticites* genus, represented mainly by *Pr. jucundus* Rem. These forms associate with less abundant *Obsoletes* sp. (Samples 1–7); however, the section is lacking species of the *Montiparus* genus in the proper beds. Similar distribution patterns are characteristic of the Podcherem-47 section in the Morchanov-Kyrta area and also of the Zaostrennaya River locality (Fig. 3), where the *Montiparus* Zone corresponds to the interval of missing zonal taxa. This feature of fusulinid distribution may point to a hiatus in sedimentation.

In the Zaostrennaya River locality of the Chernyshev Range, we identified the *Triticites quasiarcticus–T. acutus* Zone (Exposure 6, Sample 222) not by the index species occurrence, but on the basis of associated species. These were *Triticites schwageriniformis* Raus. et Bel., *T. schwageriniformis mosquensis* Ros., *T. petschoricus* Raus. et Bel., *T. petschoricus brevis* Raus. et Bel., and *T. whitei* Raus. et Bel. recovered from the member of hydractinoid limestones of the section.

A similar assemblage of fusulinids is distinguished in limestones of the same type in nearby Exposure 7 (Samples 209, 495, and 960b). *Triticites acutus* Dunb. et Cond. representing one index species of the aforementioned zone was discovered in upper beds of the exposure (Samples 216, 501, and 963, Fig. 4) together with *Triticites whitei* Raus. et Bel., *T. simplex* (Schellw.), *T. subacutus* Z. Mikh., *T. kuibyshevi* Raus., *T. noinskyi* Raus., *T. schwageriniformis* Raus. et Bel., *T. schwageriniformis mosquensis* Ros., *T. schwageriniformis nanus* Ros., *Paleostaffella moelleri* (Ozawa), and *Reitlingerina timanica* (Raus.).

In the southern part of the Chernyshev Range (Bol'shaya Sar'yuga River), dominant species of the *Triticites quasiarcticus–T. acutus* Zone are represented by *Triticites petschoricus* Raus. et Bel., *T. whitei* Raus. et Bel., and *T. schwageriniformis* Raus. et Bel.

The fusulinid zonation is also established in the Kasimovian deposits of the Verkhnyaya Pechora River basin, where Exposure 58 (Fig. 5) consists of the following units:

(a) The *Protriticites* Zone, whose index species is discovered only in Sample 790; the zone also includes *Fusulinella schwagerinoides* Deprat (samples 108 and 115);

(b) The *Montiparus montiparus* Zone with *M.* cf. *montiparus* (Ehr.), *M. umbonoplicatus* (Raus. et Bel.), and *M. subcrassulus* (Ros.) (Samples 123 and 125);

Fig. 2. Stratigraphic ranges of most important foraminifer species of the Late Carboniferous–Early Permian, the Shar'ya-80 reference section (Figs. 2–5): 1, *Schubertella sphaerica* Sul.; 2, *Sch. paramelonica* Sul.; 3, *Sch. transitoria* Sul.; 4, *Fusulinella bocki* Moell.; 5, *F. schwagerinoides* Dep.; 6, *Pulchrella pulchra* (Raus. et Bel.); 7, *Usvaella usvae* (Dutk.); 8, *Protriticites subschwagerinoides* Ros.; 9, *P. pseudomontiparus* Ros.; 10, *P. plicatus* Kir.; 11, *P. globosus* Put.; 12, *P. ovatus* Put.; 13, *Obsoletes biconicus* Kir.; 14, *O. obsoletus* (Schellw.); 15, *Montiparus montiparus* (Moell.); 16, *M. paramontiparus* (Ros.); 17, *M. umbonoplicatus* (Raus. et Bel.); 18, *M. subcrassulus* (Ros.); 19, *M. tenebrosus* (Z. Mikh.); 20, *M. sinuosus* (Ros.); 21, *M. sinuosus boreus* Z. Mikh.; 22, *Triticites petschoricus* Raus. et Bel.; 23, *T. petschoricus varsanofievae* Z. Mikh.; 24, *T. whitei* Raus. et Bel.; 25, *T. quasiarcticus* Sol.; 26, *T. acutus* Dunb. et Cond.; 27, *T. subacutus* Z. Mikh.; 28, *T. simplex* Schellw.; 29, *T. noinskyi* Raus. et Bel.; 30, *T. noinskyi plicatus* Ros.; 31, *T. variabilis* Ros.; 32, *T. secalicus* Say; 33, *T. fortissimus* Raus.; 34, *T. schwageriniformis* Raus.; 35, *T. schwageriniformis nanus* Ros.; 36, *T. schwageriniformis mosquensis* Ros.; 37, *T. exilis* Pant.; *38, T. shikhanensis compactus* Ros.; 39, *T. parvulus* Schellw.; 40, *T. kuibyshevi* Raus.; 41, *T. rossicus* Schellw.; 42, *T. rossicus gzhelicus* Bensh; 43, *T. scharjuensis* Z. Mikh.; 44, *T. pseudoarcticus* Raus.; 45, *T. paraarcticus* Raus.; 46, *T. stuckenbergi* Raus.; 47, *T. ventricosus* (Meek); 48, *T. elongatissimus* Ros.; 49, *T. subschwagerinoides* Grozd.; 50, *T. uniensis* Grozd.; 51, *T. obuncus* Konov.; 52, *T. opinabilis* Konov.; 53, *T. ordinarius* Konov.; 54, *T. adjunctus* Konov.; 55, *Jigulites jigulensis* Raus.; 56, *J. volgensis* (Raus.); 57, *J. longus* (Ros.); 58, *J. longus formosus* (Ros.); 59, *Daixina tschernovi* Z. Mikh.; 60, *D. privilegiata* (Pant.); 61, *D. sokensis* Raus.; 62, *D. robusta* Raus.; 63, *D. vozhgalensis* Raus.; 64, *Rugosofusulina prisca* (Ehrenb.); 65, *R. prisca uralica* Z. Mikh.; 66, *R. aktjubensis* Raus.; 67, *R. burkemensis* Vol.; 68, *Schwagerina fusiformis* Krot.; 69, *Sch. vulgaris* Scherb.; 70, *Sch. vulgaris aktjubensis* Scherb.; 71, *Sch. vulgaris timanica* Scherb.; 72, *Sch. vulgaris ashensis* Scherb.; 73, *Sch. moelleri* Raus.; 74, *Sch. pavlovi* Raus.; 75, *Sch. borealis* Scherb.; 76, *Sch. shamovi* Scherb.; 77, *Pseudofusulina paragregaria* Raus.; 78, *P. gregaria* Lee; 79, *P. paragregaria simplex* Z. Mikh.; 80, *P. anderssoni* (Schellw.); 81, *P. anderssoni elongata* Scherb.; 82, *P. eliseevi* Z. Mikh.; 83, *P. fecunda* Sham. et Scherb.; 84, *P. parafecunda* Sham. et Scherb.; 85, *P. rhomboides* Sham. et Scherb.; 86, *P. paradoxa* Sham. et Scherb.; 87, *P. venusta* Konov.; 88, *P. lutuginiformis* Raus.; 89, *P. sulcata* Korzh.; 90, *P. paragregaria ascedens* Raus.; 91, *Globifusulina nux* (Schellw.); 92, *G. globulus* (Raus.); 93, G. *sphaeroidea* (Raus). Limestone (1–9): detrital, mainly thick-bedded (1), detrital thin-bedded (2), sandy (3), clayey (4), hydractinoid (5), crinoidal (6), bryozoan (7), fusulinid (8), coral (9). Calcareous gravelstone, or conglomerate (10); calcareous breccia (11); cherty concretions and interlayers (12); sandstone (13); siltstone (14).

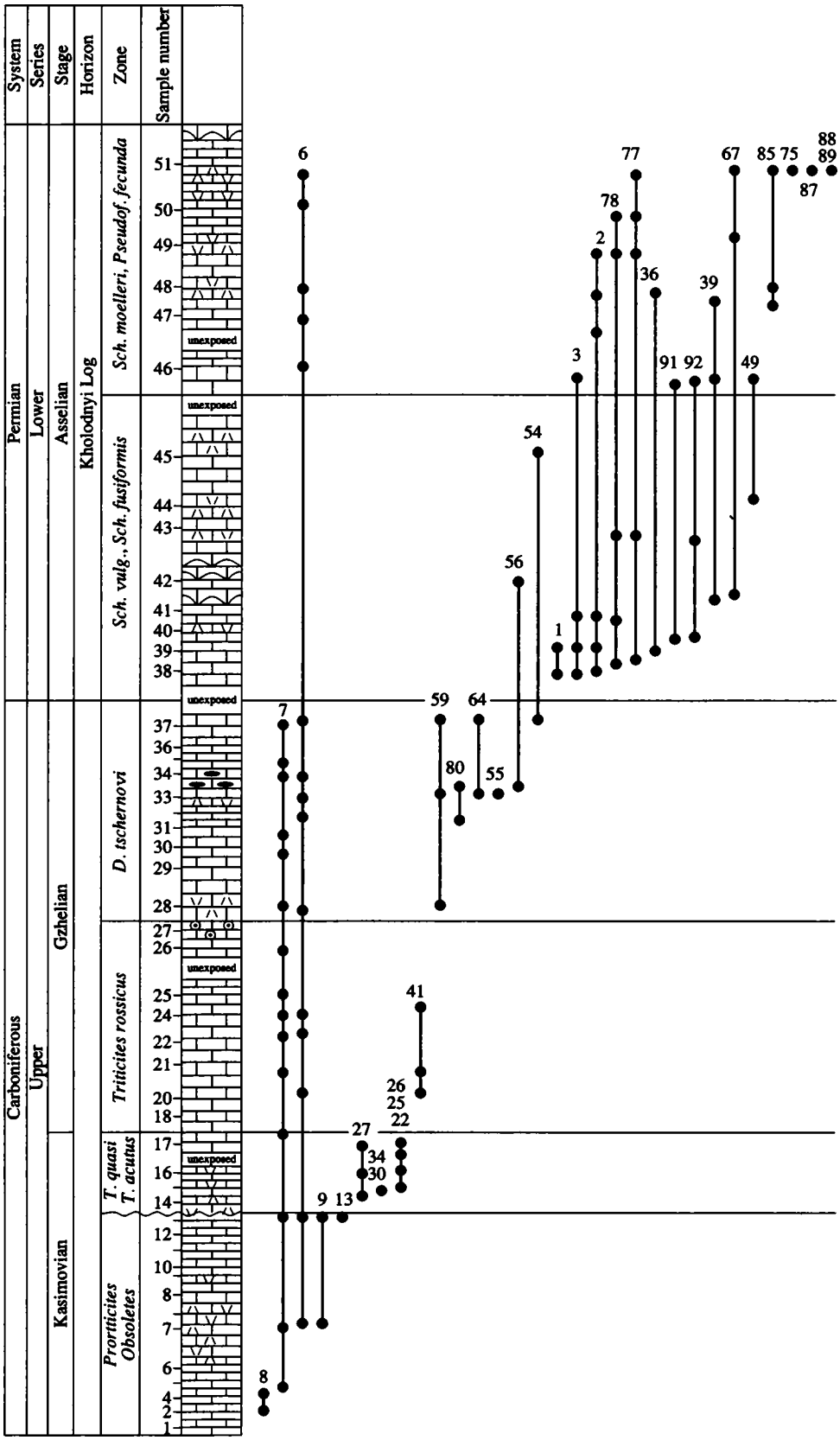


Fig. 3. Stratigraphic ranges of the most important foraminifer species of the Late Carboniferous–Early Permian, the Podcherem River, Exposure Morchanov-Kyrta-47 (list of species and other symbols same as in Fig. 2).

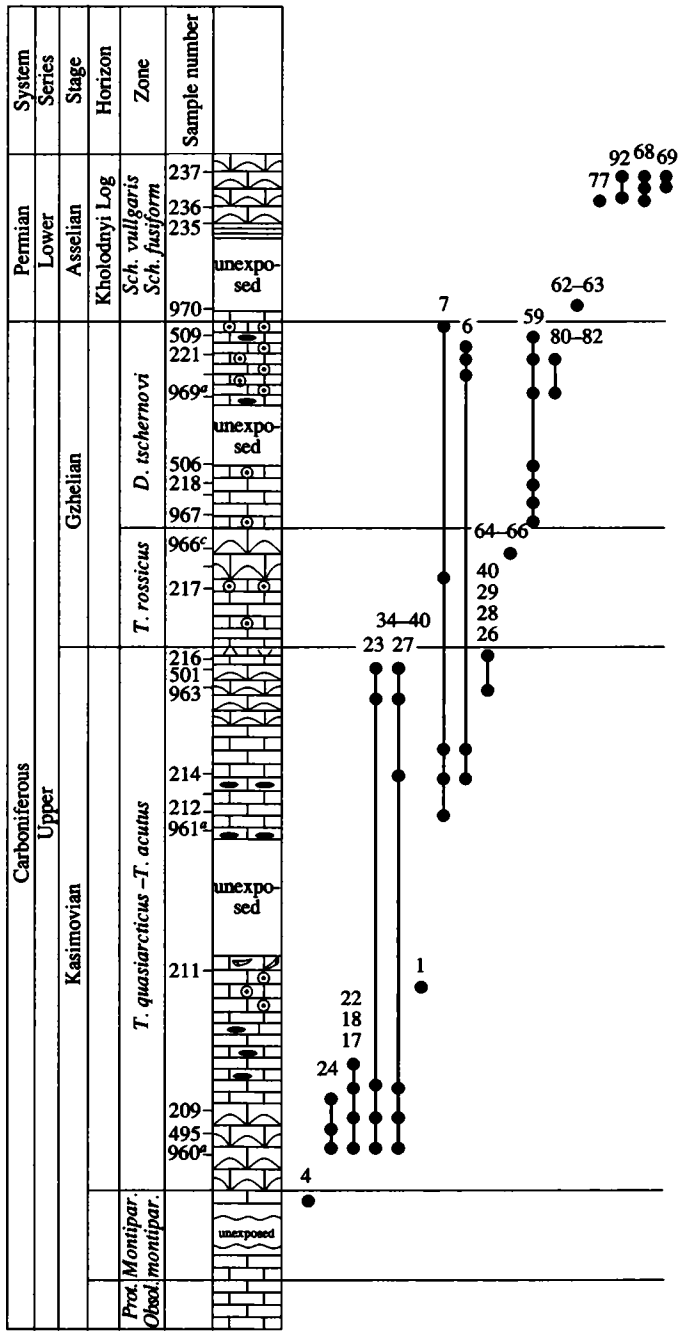


Fig. 4. Stratigraphic ranges of the most important foraminifer species of the Late Carboniferous–Early Permian, the Zaostrennaya River, exposures 7 and 8 (list of species and other symbols same as in Fig. 2).

(c) The *Triticites quasiarcticus–T. acutus* Zone is distinguished by the occurrence level of its first index species (Samples 774, 152, 153, and 772); above this level, it also includes *T. acutus* Dunb. et Cond. (Sample 154).

In other sections composed mainly of hydractinoid carbonates (Figs. 2–5), the Kasimovian Stage is either undetected or represented by its upper undivided interval.

Zonal taxa of the Gzhelian Stage. As we noted in the introduction, Gzhelian deposits of the region comprise the *Triticites rossicus* Beds and *Daixina sokensis* Zone. Recently, we reported (Chermnykh and Mikhailova, 1991) that the first taxon is scarce here and often replaced in its vertical and horizontal distribution ranges by *Daixina tschernovi* Z. Mikh.

The lower Gzhelian *Triticites rossicus* Beds are well represented in the reference section Shar'yu-80. Their index species occurs here beginning from the level of Sample 159 together with *Triticites rossicus gzhelicus*

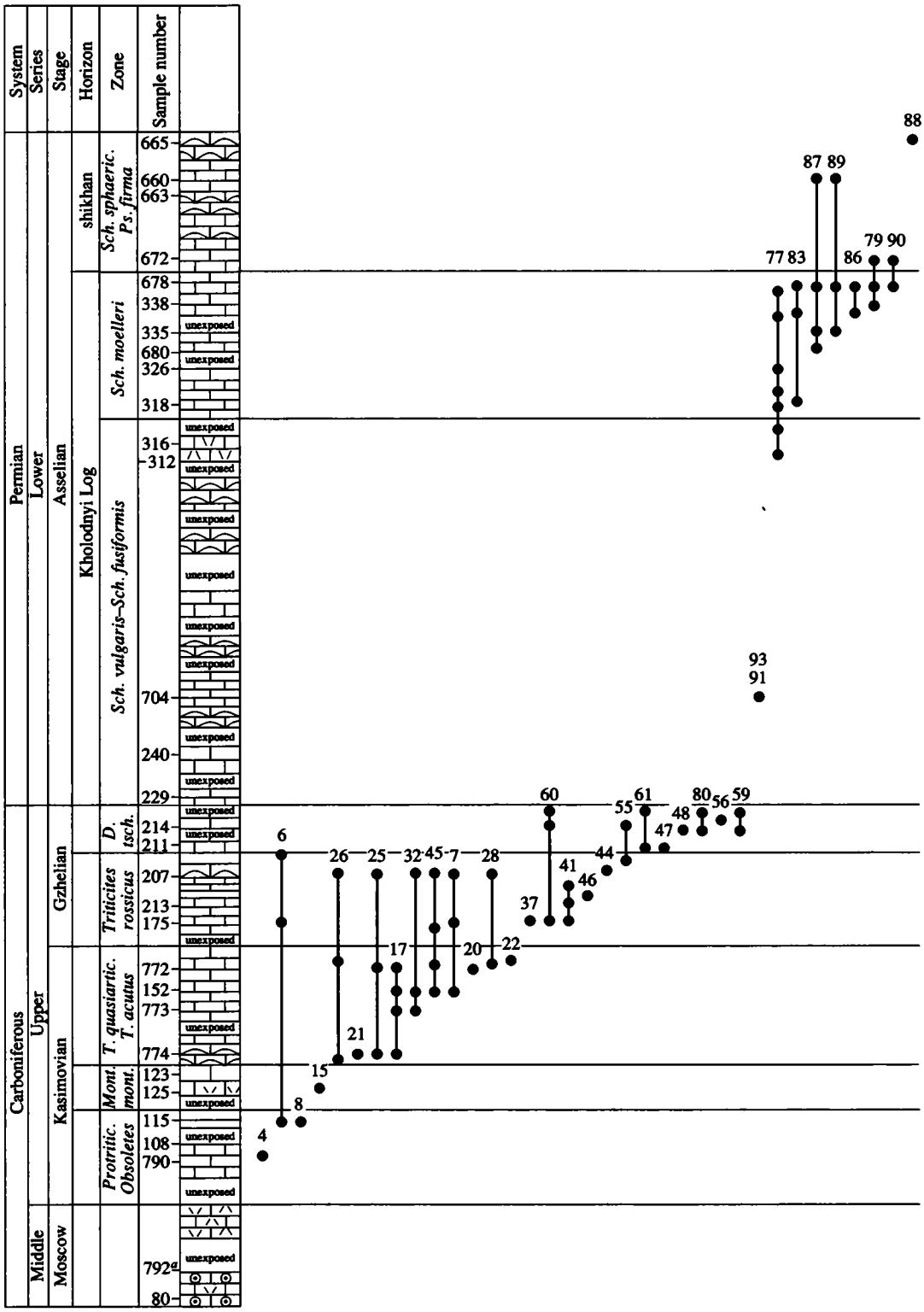


Fig. 5. Stratigraphic ranges of the most important foraminifer species of the Late Carboniferous–Early Permian, the Pechora River, Exposure 58 (list of species and other symbols same as in Fig. 2).

Bensh, *T. secalicus* Say, *T. simplex* Schellw., *T. scharjuensis* Z. Mikh., and *T.* cf. *pseudoarcticus*. Above this level, we identified *Triticites rossicus* forms in Samples 163, 23, 23a, 165, 24, 24a, and 857.

This section demonstrates the unusually late appearance of *Jigulites* forms. The first level of their occurrence corresponds to Sample 170, in which *Jigulites longus* (Ros.) and *J. longus formosus* (Ros.) coexist

with the index species *Daixina tschernovi* Z. Mikh. of the younger zone.

In the section Verkhnyaya Pechora-58, *Triticites rossicus* is rather abundant (Samples 175–178, 762–764, and 736); however, we recorded its coexistence with *Triticites stuckenbergi* Raus. (Fig. 5) only in the upper part of the respective zone. Species of the index genus *Jigulites* are missing in this unit, and their first occurrence level is confined to basal beds of the overlying *Daixina* Zone. All the data we have allow us to suggest that *Daixina tschernovi*, the index species of the last zone, is terminal in the fusulinid succession of the Gzhelian Stage. It is widespread and reported from many sections: Shar'yu-80, Zaostrennaya-6, Zaostrennaya-7, Podcherem-49, Verkhnyaya Pechora-58, Un'ya-28, Bol'shaya Shaitanovka, Ilych, Ul'vozh, and Gerd'yu. However, only the Bol'shaya Pechora-58 and Un'ya sections shown among others in Figs. 2–5 yielded another index species, *Daixina sokensis*. In addition, the last form was reported in the Gerd'yu, Bol'shaya Shaitanovka, and Ilych localities (Fig. 1).

The Bol'shaya Shaitanovka and Un'ya sections also contain abundant *Jigulites volgensis* (Raus.), *J. longus longus* Ros., *L. longus formosus* Ros., *J. Longus mucronatus* Ros., *J. magnus* Ros., *Daixina scherbovichae* Z. Mikh., and *D. privilegiata* Pant.

As already mentioned, we noted the lateral replacement of index species *Daixina sokensis* by *Daixina tschernovi* and also the fact that these vicariant forms coexist with fusulinid assemblages similar in composition. This observation suggests that both species are coeval.

The *Daixina* zones of higher levels are recognized in the Chernyshev Range (the Shar'yu and Zaostrennaya rivers) and at the western slope of the Urals (the Un'ya and Gerd'yu rivers).

The *Daixina robusta* Zone of the Chernyshev Range, which also yielded *Daixina vozhgalensis* Raus., successively overlies the *Daixina tschernovi* Zone in the section. The first occurrence level of schwagerinids in this section is above the stratigraphic range of *Daixina robusta*, and similar distribution patterns are characteristic of the Gerd'yu, Un'ya, and Ilych (Moleben-El') sections.

The only case when zonal species *Daixina robusta* Raus. were found to coexist with *Schwagerina vulgaris* Scherb. is Exposure 28 at the Un'ya River, in a rock outcrop separated from others by a considerable unexposed interval.

According to the standard schemes, the first occurrence level of *Daixina robusta* marks basal beds of the Permian System. However, this species rarely occurs together with *Schwagerina* forms. For instance, in Exposure 4 at the Gerd'yu River, we distinguished *Daixina sokensis* Raus., *D. vozhgalensis* Raus., and *D. robusta* Raus. in Bed 31 and also found that the last form is typical of overlying Bed 32. The first *Schwagerina* species, namely *Schwagerina vulgaris* Scherb., appears here only one and a half meters above the last bed (Chermnykh and Mikhailova, 1991). Accordingly, we believe that *D. robusta* Beds represent the lower subzone or horizon of the *Schwagerina vulgaris–Sch. fusiformis* Zone, and the Carboniferous–Permian boundary should be placed at the base of this subzone.

DISCUSSION

Our revision of reference Upper Carboniferous sections consisting mainly of shallow-water detrital deposits with abundant hydractinoid bioherms in both the Kasimovian and the Gzhelian intervals (the maximum abundance of such bioherms is confined to the Asselian Stage of the Lower Permian) revealed the irregular distribution of zonal taxa in the region studied.

Communities of the Kasimovian Stage. Index species of the *Protriticites pseudomontiparus–Obsoletes obsoletus* Zone are usually disjoined. Their coexistence is recorded only in the Shar'yu-80 section (Fig. 2) and not encountered in many other localities, such as the northern Novaya Zemlya and Pai-Khoi (Chaika Cape) regions and the Kos'yu, Kozhim, Shchugor, Ilych, and Bol'shaya Shaitanovka river basins. As shown above, the first index species of this zone was found in the Nizhnie and Verkhnie Vorota localities of the Shchugor River, in outcrops of the Kos'yu River, and in the Northern Island (Litke Cape) of the Novaya Zemlya Archipelago. In other studied localities, the same stratigraphic level also yielded other representatives of the genus *Protriticites*. For instance, this level in the Shar'yu reference section was found to bear *Protriticites pseudomontiparus* Put., *Pr. plicatus* Kir., *Pr. globulus* Put., *Pr. ovatus* Put., and *Pr. ovoides* Put. In some localities lacking the index species, we detected *Pr. rotundus* Grozd. (the Verkhnie Vorota locality at the Shchugor River) and *Pr. jucundus* Rem. (the Podcherem-49 section in the Oselka-Kyrta area).

Likewise, irregular distribution is characteristic of the index species *Obsoletes obsoletus*, which we encountered only in the Verkhnie Vorota, Shar'yu, and Kos'yu localities.

The next index species *Montiparus montiparus* is even less widespread, because the respective interval in many sections either corresponds to the hiatus in sedimentation or is unexposed. For instance, in the Shchugor-12 section (Nizhnie Vorota locality), beds bearing *Protriticites globosus* Put., *Pr. pseudomontiparus* Put., and *Quasifusulina longissima* Moell. are immediately overlain by sandy limestone with *Triticites quasiarcticus* Sol. and *T. acutus* Dunb. et Cond. (Chermnykh and Mikhailova, 1970). The same transgressive overlapping of the *Triticites quasiarcticus–T. acutus* Zone was observed in the Podcherem-47 section and may be present in two localities (6 and 7) at the Zaostrennaya River. Deposits of this zone compose the uppermost interval of the Kasimovian Stage and are widespread in the regions under consideration. In addi-

tion to the localities indicated in Figs. 2–5, we encountered *Triticites quasiarcticus* in the Ul'vozh, Bol'shaya Sar'yuga, Kos'yu, Inta, Kozhim, Gerd'yu, Ilych, and Ultan river basins. However, *Triticites acutus* has a somewhat narrower distribution range.

Communities of the Gzhelian Stage. As shown in the previous section, the Gzhelian Stage of the North Uralian region has the two-member structure, in contrast to its stratotype. The *Triticites rossicus* Beds described earlier by Mikhailova (1974) are suggested to represent its lower unit. The index species of the beds, including subspecies *T. rossicus rossicus* Schellw. and *T. rossicus gzhelicus* Bensh, occur in all sections of the Chernyshev Range, where this part of the Upper Carboniferous is exposed, and also in outcrops of the Kozhim and Ilych rivers. The peculiar feature of this unit consists in the nearly complete absence of another index species, such as *Triticites stuckenbergi*, whose coexistence with *T. rossicus* was recorded only in the Shar'yu, Pechora, and Ilych sections.

In the stratotype area of the stage, *T. rossicus* coexists with *T. paraarcticus*, and both mark the local zone corresponding to the Rusavkino Horizon (*Putevoditel'* ..., 1975). The *Triticites stuckenbergi* Zone here is immediately above the latter and corresponds to the Amerovo Horizon (Well 7, the left bank of the Kirzhach River). Unfortunately, in the accepted stratigraphic scheme of the Upper Carboniferous of the Russian platform, these two well-detectable units are coupled into one *T. rossicus–T. stuckenbergi* Zone (*Reshenie* ..., 1990).

In the stratotype, these units are overlain by beds with *Jigulites jigulensis* and *J. longus*, and then by *Daixina sokensis* and *Pseudofusulina gregaria* zones. Thus, the sections we studied differ from the stratotype in that they lack clear stratigraphic levels of *T. stuckenbergi* Raus.

The next difference between the Gzhelian fusulinids from the stratotype area and North Uralian sections is in the abundance of *Daixina* forms in the last region, where these species are confined to beds overlying the *Triticites rossicus* Zone. As we reported earlier (Chermnykh and Mikhailova, 1991), especially abundant in our region are specimens of *Daixina tschernovi*, which enabled us to interpret the respective zone as correlative with the *D. sokensis* Zone.

Localities Zaostrennaya-6 and Zaostrennaya-7 are the most important for detecting successive changes in the Gzhelian fusulinid communities in the North Uralian region. In the first locality, we detected the first occurrence level of index subspecies *D. tschernovi tschernovi* and *D. tschernovi regularis* in Sample 225 (Fig. 4). These index taxa were also distinguished in Samples 226, 230, and 231, and in the last case they coexist with *Pseudofusulina eliseevi* Z. Mikh. ex gr. *Pseudofusulina anderssoni* and *Schubertella paramelonica minor* Sul. In Sample 232, we identified *Jigulites volgensis* Raus., *Pseudofusulina anderssoni* (Schellw.), *P. eliseevi* Z. Mikh., and *Schubertella paramelonica* Sul. Sample 233 from the higher level yielded the same assemblage and also *Pseudofusulina anderssoni anderssoni* (Schellw.) and *P. anderssoni elongata* Scherb.

Higher in the section, after a gap in exposures of about 10 m, an outcrop of limestone yielded *Daixina vozhgalensis* Raus., *Triticites parvulus* Schellw., *T. schwageriniformis nanus* Ros., *Schubertella sphaerica* Sul., *Sch. paramelonica minor* Sul., and some other forms characteristic of the Asselian Stage of the Lower Permian. *Schwagerina* species are missing at this level. Similar distribution patterns (fossiliferous Samples 506, 967, 969, 969A, and 218) were observed in the Zaostrennaya-7 section, where, as in the previous locality, we detected the *Pseudofusulina anderssoni–P. eliseevi* Beds (Samples 509, 969, 220, and 221) above the occurrence level of *D. tschernovi* Z. Mikh. The Lower Permian deposits are distinguished here on the basis of discovered *Daixina vozhgalensis* Raus. and *D. robusta* Raus., which belong to the lower subzone of the *Schwagerina fusiformis–Sch. vulgaris* Zone (Sample 970). The first occurrence level of both *Schwagerina* forms is several meters higher, and here they coexist with *Pseudofusulina paragregaria* (Samples 972, 973, 235, and 236). This coexistence suggests that, in sections lacking *Schwagerina* forms, the lower boundary of the Asselian Stage can be placed at the first occurrence level of *Pseudofusulina paragregaria* Raus.

Sections investigated at the Zaostrennaya River (the Chernyshev Range) show that bed-by-bed studies of the Carboniferous–Permian boundary deposits can help to reconstruct the successive changes in respective fusulinid communities. The fusulinid succession was presumably as follows: *Daixina tschernovi–Pseudofusulina anderssoni, P. eliseevi–Daixina robusta, D. vozhgalensis–Schwagerina fusiformis, Sch. vulgaris, Pseudofusulina paragregaria*. Other ideas to regard the *Daixina tschernovi* Beds as correlative with the *Daixina bosbytaensis–D. robusta* Zone (Davydov *et al.*, 1994), or to consider them as representative of the lower unit of the Asselian Stage (Konovalova, 1991), seem poorly substantiated, if we consider the following data.

Analyzing data on wells drilled in the Pechora syneclise, we were concerned with the data of Konovalova (1991, pp. 57–58), who presented the following data on the Well-102 section in the Vozei area:

(1) The depth interval of 1648–1633 m corresponds to the *Pseudofusulina narjanmarica* Beds bearing this index species in association with *Ps. volozhaninae* Konov., *Rugosofusulina praevica agregia* Schlyk., *Triticites aijuvensis* Konov., and *Usvaella usvae* (Dutk.);

(2) The depth interval of 1633–1626 m represents the *Daixina tschernovi* Beds with *D. tschernovi tschernovi* Z. Mikh., *D. tschernovi regularis* Z. Mikh., and *Triticites adjunctus* Konov.;

(3) The depth interval of 1626–1602 m comprises beds with new small *Triticites* and *Daixina* forms, such as *T. uberatis* Konov., *T. lautus* Konov., *D. vozeica*

Konov., *D. sedujachaensis* Konov., *D. grata* Konov., *D. lilia* Konov., and *D. robusta mendosa* Konov., associated with *Pseudofusulina paragregaria* Raus., *Ps. paragregaria* f. *simplex*. Z. Mikh., *Ps.* ex gr. *gregaria* Lee., and *Ps.* ex gr. *anderssoni* (Schellw.).

As one can see, the uppermost Member 3 (or beds with "new small *Triticites* and *Daixina* forms") is of the Early Permian age, because it yielded *Daixina robusta mendosa* Konov. and *Pseudofusulina paragregaria* Raus.

Beds underlying it (Member 2) cannot be attributed to the Permian System, because they contain the Late Carboniferous species *Daixina tschernovi* Z. Mikh. spread all over the western slope of the Urals, whereas zonal species of Member 3 are missing here.

Describing the section she studied, Konovalova reported (*ibid.*, p. 59) that *Daixina robusta mendosa* Konov. occurs "directly at the top of the *Daixina tschernovi* Beds," and that *Pseudofusulina paragregaria* together with *Ps. anderssoni* are recorded "below *D. tschernovi* Beds." However, on page 58, she presented another and correct succession of fusulinid assemblages recovered from Well 102 in the Vozei area.

A comparable fusulinid succession was also recovered from Well Naryan-Mar-1. In this case, samples from the depth interval of 1866–1882 m yielded *Triticites* aff. *rossicus* (Schellw.) and *Jigulites jigulensis* Raus., whereas *Daixina tschernovi* Z. Mikh. and *Pseudofusulina narjanmarica* Konov. were detected higher in the section (depth interval of 1845.5–1866 m).

Pseudofusulina paragregaria Raus. and *Schubertella* forms were recovered from the depth interval of 1801.7–1809.7 m, and *Pseudofusulina paragregaria ascedens* Raus., *Daixina kachgortica* Konov., and again *Schubertella* species were encountered in the depth interval of 1772–1797.7 m.

According to these fusulinid assemblages, only two upper intervals are compatible in our opinion with the Asselian Stage.

Similar successions of zonal fusulinid species can be also illustrated by data on many other drilling localities: Well 1 in the Kochmes area; wells 1, 47, and 61 in the Khar'yaga area; Well 23 in the Srednii Vuktyl area; and wells 4, 6, and 13 in the Usa–Kolva area.

To complete our discussion on sections recovered by drilling in the Pechora syneclise, we should note that the core sampling was fragmentary in all the wells drilled here, and this hindered the correct interpretation of the data obtained. We do not exclude the possible coexistence of *Daixina tschernovi* Z. Mikh. with definitely Early Permian index species. Their association seems quite possible for the uppermost portion of the *D. tschernovi* stratigraphic range, where the younger zonal taxon could be of key significance; however, even this situation is compatible with our understanding of the biostratigraphic zone.

Refining zonal subdivisions of the Upper Carboniferous in the North Uralian region, we found them dissimilar with horizons distinguished in the stratotype sections of the southern Urals (Chuvashov *et al.*, 1989).

Our *Triticites rossicus* Zone is considered as corresponding to the *Triticites stuckenbergi* Zone, because we discovered the last index species in association with *T. rossicus*, and this interval, being coupled with the lower part of the *Daixina tschernovi* Zone, seemingly represents a unit correlative with the Azantash Horizon from the scheme of Chuvashov and his colleagues. Their Martuk Horizon can be correlated with the upper part of the North Uralian *Daixina tschernovi* Zone.

The comparison of our zonal subdivisions with the standard Upper Carboniferous horizons of the Russian platform (*Reshenie ...*, 1990) shows that the *Triticites rossicus* Zone initially recognized as synonymous beds corresponds to the Dobryatino Horizon, whereas the *Daixina tschernovi* Zone overlying this unit seems to be an analogue of the Pavlovskii Posad and Noginsk horizons.

CONCLUSION

The Upper Carboniferous sections of the North Uralian region display the following specific features.

The Kasimovian Stage here includes three fusulinid zones, whose deposits are widespread in the Urals and Russian platform. These are the *Protriticites pseudomontiparus–Obsoletes obsoletus, Montiparus montiparus*, and *Triticites quasiarcticus–T. acutus* zones.

The Gzhelian Stage of the region is of the two-member structure and comprises the *Triticites rossicus* and *Daixina tschernovi* zones. The first of these zones incorporates the *Triticites stuckenbergi* Zone, and recently the respective unit of the Russian platform was named the *Triticites rossicus–T. stuckenbergi* Zone (Alekseeva *et al.*, 1984). Our second zone, whose deposits in the North Uralian region frequently contain *Jigulites jigulensis* (Raus.) and *J. volgensis* (Raus.), has a stratigraphic range equal, in our opinion, to the coupled *J. jigulensis* and *Daixina sokensis* zones accepted in the standard stratigraphic schemes of the Urals and Russian platform.

Evolution of foraminifer communities at the Carboniferous–Permian boundary is manifested in the region of our study by the transition from the Late Carboniferous *Daixina tschernovi–Pseudofusulina anderssoni* and *P. eliseevi* assemblages to the Early Permian assemblage of *Daixina robusta, D. vozhgalensis, Schwagerina fusiformis, Sch. vulgaris*, and *Pseudofusulina paragregaria*.

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Stratigraphy and Formation Environments of the Upper Permian–Triassic Sequence in the Aksu-Kendyrli Flexure Area (Southern Mangyshlak)

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Abstract—Based on complex and voluminous geologic–geophysical data, relationships of pre-Jurassic sedimentary sections in the southern flank of the Southern Mangyshlak depression are outlined. Regional unconformities were established in the section, the main one of which (pre-late Olenekian) divides Upper Permian–Triassic deposits into two structural stages. It is shown that the Aksy-Kendyrli flexure comprises three carbonate members of different age (Early, Middle, and Late Triassic), every one of which is of interest for oil and gas prospecting. Description and correlation of local stratigraphic units are given and their formation environments are characterized.

Key words: *stratigraphy, Triassic deposits, Southern Mangyshlak, seismostratigraphic analysis, sedimentation environments.*

The Aksu-Kendyrli tectonic terrace in the Triassic structure occupies an intermediate position between the Zhazgurli (Southern Mangyshlak) depression and the Karabogaz Paleozoic massif (Fig. 1). Upper Permian–Triassic deposits referred to the preplatform (transitional) structural–tectonic complex (STC) are represented by terrigenous–carbonate sequences composed of rocks of variable genesis.

For a long time data on their structure were scarce, because deep drilling was mainly conducted in the western part of the region in question (Temir-Baba, Aksu-Kendyrli, Southern Aksu). Recently, drilling materials on eastern (Zharty and Araburun) areas were studied and, thus, geological–geophysical data were sufficiently supplemented. This allowed a more detailed subdivision and correlation of the Upper Permian–Triassic sections in the Aksu-Kendyrli flexure to be performed and their age to be substantiated in some cases. Lithologic properties, marker beds, and unconformities, as well as fossils, were used when interpreting the log diagrams (Fig. 2). Simultaneously, interpretation of CMP records correlated with deep holes was performed.

Based on the seismostratigraphic analysis, two seismic–geologic sequences (SGS) were established in the transitional rock complex of the Aksu-Kendyrli flexure. They are divided by regional unconformity (reflecting horizon V_3^2) related to the pre-late Olenekian folding phase (Volozh, 1991). The lower SGS is traced in the northern and central parts of the flexure, where it comprises Upper Permian–lower Olenekian deposits. The deposits penetrated by boreholes Ataburun P-1 and Temir-Baba P-1 are strongly dislocated and fill the narrow and deep (up to 4 km) tafrogenic trough (Fig. 3). Their more complete section, in which we define two units, was penetrated by the former hole. The lower, mainly red-colored unit (Borehole Ataburun P-1, interval of 4360–4600 m; Borehole Temir-Baba P-1, interval of 4257–4503 m), which we correlate with the Amanbulak Formation of Tuarkyr, consists of sandstones and tuffaceous sandstones interbedding with shales, gritstones, and conglomerates. Sandstones and tuffaceous sandstones are polymictic, coarse-grained, highly hydromicaceous, and bear clasts of vitroclastic tuffs. Gritstones and conglomerates are composed of fragments of quartzite and porphyry (up to 60%); quartz and siliceous rocks with sericite make up less than 30%. Shales are carbonate, with obscure horizontal bedding and with carbonate nests and pyrite concretions. Organic remains were not found here; therefore, the Amanbulak Formation is conditionally dated as the Late Permian–Early Triassic according to its stratigraphic position. The upper, mainly variegated, member is present only in the section recovered by Borehole Ataburun P-1 (interval of 3785–4360 m). It consists of shales and siltstones with sandstone interbeds. A.Yu. Lopato identified conchastracans *Lioestheria mokrinskyi* Mol. and *Pseudoestheria tumariana* Nov. from the interval of 3890–3900 m. Based on these finds

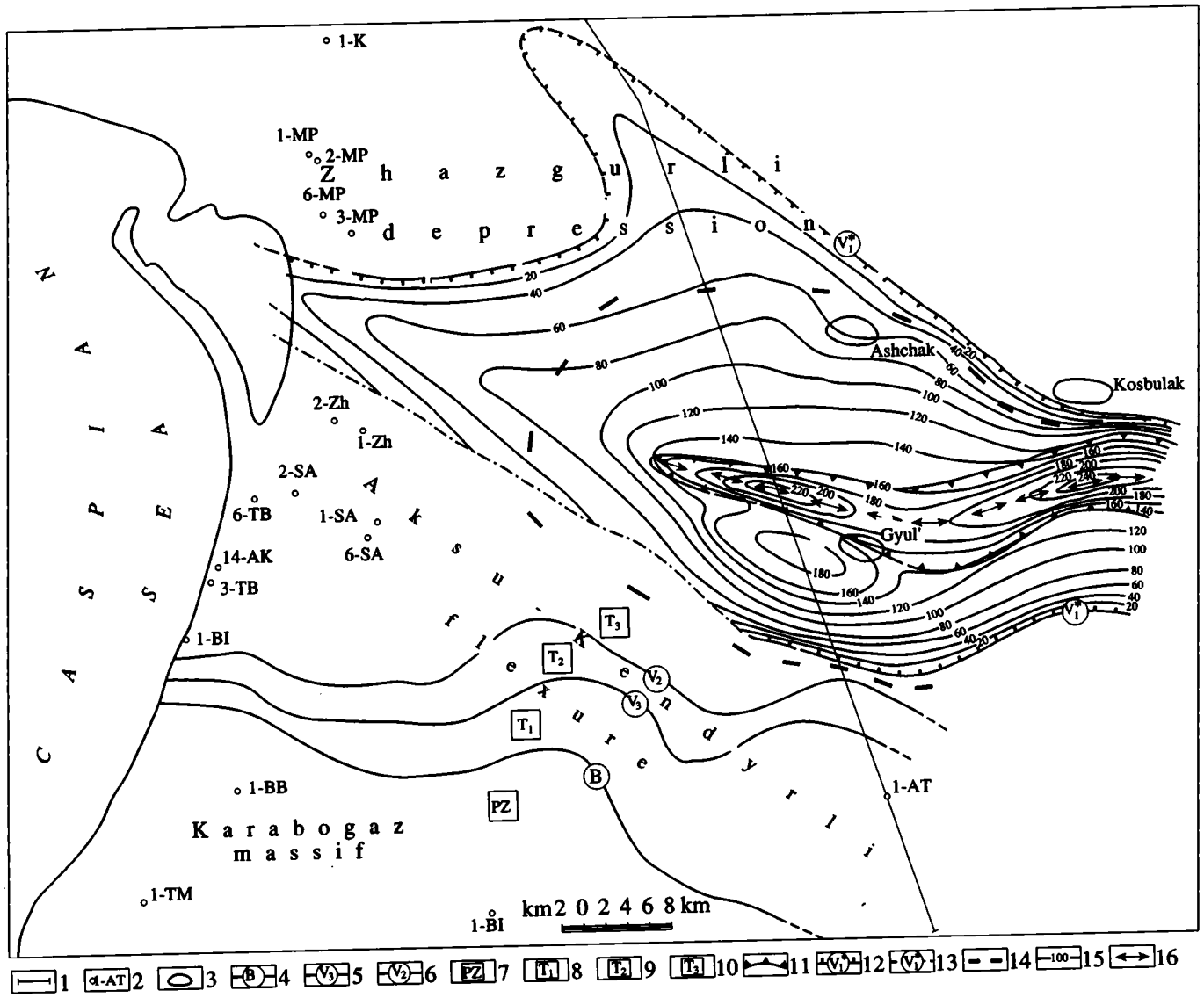

Fig. 1. Geologic map of the Pre-Jurassic erosional surface in the southern flank of the Southern-Mangyshlak depression. 1, Location of the regional profile (Fig. 4); 2, boreholes: SA, Southern Aksu; TB, Temir-Baba; AK, Aksu-Kendyrli; K, Kaundy; BB, Bukbash; BI, Birbas; TM, Tamdy; AT, Ataburun; MP, Makat-Pribrezhnaya; ZH, Zharty; 3, local structures in the Triassic; 4–6, contours at the pre-Jurassic erosional surface of : 4, basement; 5, Lower Triassic; 6, Middle Triassic; 7–10, distribution at the pre-Jurassic erosional surface of: 7, Paleozoic shales; 8, Lower Triassic; 9, Middle Triassic; 10, Upper Triassic; 11–14, boundaries of the Upper Triassic facies: 11, filling of the Late Triassic erosional incision; 12, filling of the Late Triassic synsedimentary depression; 13, arbitrary southern boundary of depression filling (loss of correlation); 14, lateral accretion; 15, contours of equal time-thickness Δt (ms) of incision and depression infilling; 16, axial line (the deepest part) of incision.

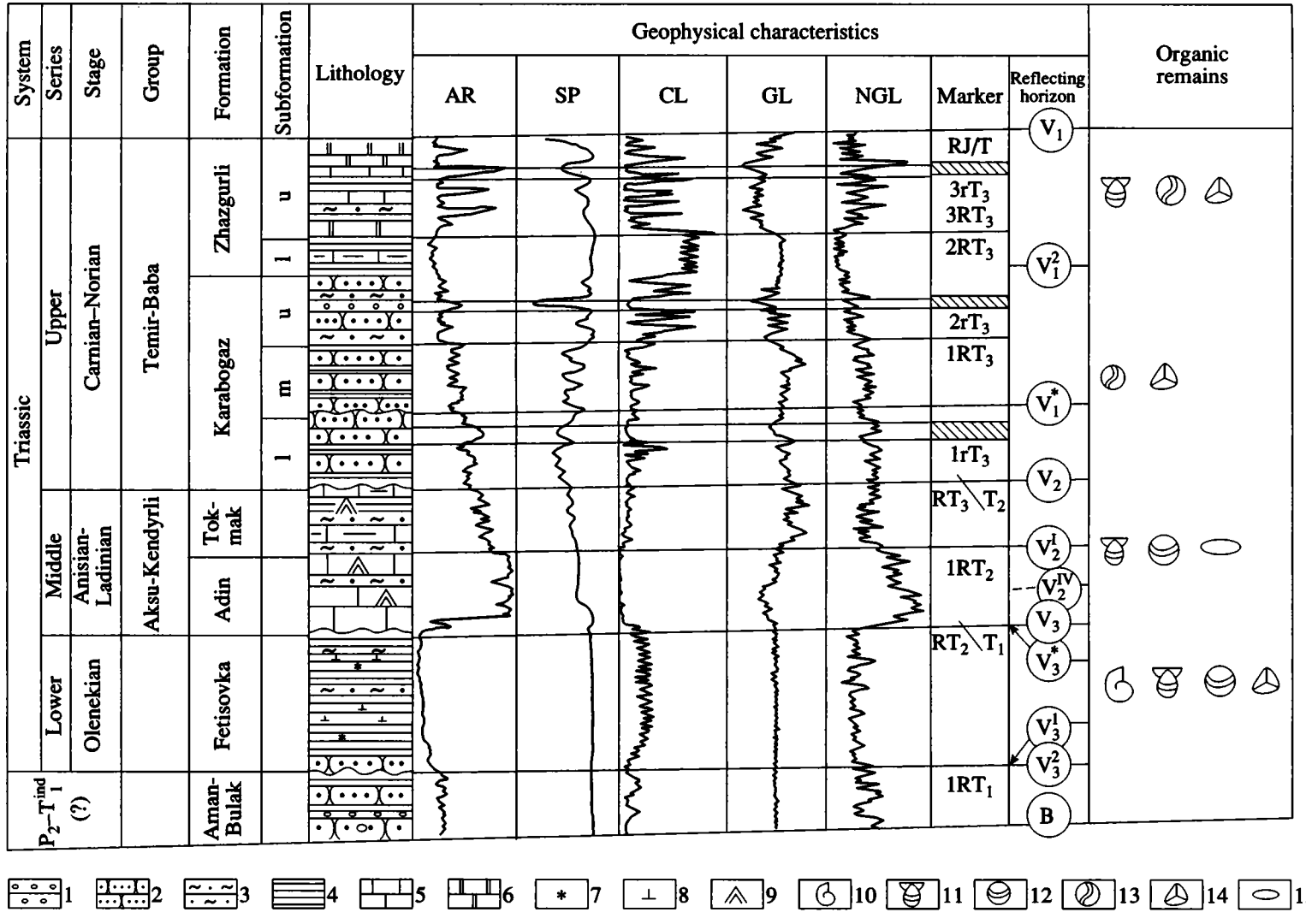

Fig. 2. Stratigraphic resolution and correlation of the Upper Permian–Triassic deposits in the Aksu-Kendyrli flexure by different methods. 1, Pebble, gravel; 2, sandstone; 3, siltstone; 4, argillite; 5, limestone; 6, dolomite; 7, red coloration; 8, carbonate admixture; 9, tuffaceous material admixture; 10–15, organic remains: 10, ammonoids; 11, bivalves; 12, conchostracans; 13, charophytes; 14, spores and pollen; 15, ostracods. AR, apparent resistivity; SP, spontaneous polarization; CL, caliper logging; GL, gamma-ray logging; NGL, neutron gamma-ray logging.

and the stratigraphic position of the unit, we refer it to the lower Olenekian Substage.

The upper SGS comprises the upper Olenekian–Upper Triassic deposits, which are transitional according to the intensity of folding between pre-upper Olenekian and Jurassic–Cretaceous deposits. Based on unconformities, this seismogeologic sequence can be subdivided into the upper Olenekian, Middle Triassic, and Upper Triassic seismic units (reflecting horizons V_3 and V_2 mark the top of the Lower and Middle Triassic deposits, respectively). Southward, all Triassic units rise toward the Karabogaz massif and are successively cut by the pre-Jurassic erosional surface (reflecting horizon V_1).

The upper Olenekian seismostratigraphic unit (SU) corresponds to the upper Olenekian deposits from 300 m to 1200 m thick, which fill the narrow trough in the Aksu-Kendyrli flexure. We divide this unit into two seismic facies. The lower one is bounded at the top by reflecting horizon V_3^1 and completely wedges out northward near the paleouplift slope (Fig. 4). This seismic facies is correlative to clayey–carbonate gray-colored deposits penetrated only by Borehole Ataburun P-1 (interval of 3495–3785 m). They are represented by shallow marine dark gray fine-crystalline limestones with a tuffaceous material admixture and, less commonly, by organogenic–detrital limestones with dark gray shale, tuffaceous shale, and tuff interbeds. Ammonoids and bivalve mollusks determined by V.A. Gavrilova unequivocally testify to the late Olenekian age of these deposits; moreover, the presence of *Dorikranites bogdoanus* (Buch) allows them to be referred to the lower upper Olenekian.

The clayey unit corresponding to the upper seismic facies consists of two members that are divided by the reflecting horizon V_3^*; the upper member is cut in the north by the pre-Late Triassic erosional surface (Figs. 4, 5). The lower member is penetrated by Borehole Ataburun P-1 (interval of 3307–3495 m). It is composed of brown, greenish gray, and gray siltstones and sandstones, whose age has not been paleontologically substantiated.

Based on the regional correlation of reflecting horizons and borehole sections, we assume that the upper member in the western part of the Aksu-Kendyrli flexure corresponds to the Fetisovka Formation. The latter is composed of shallow marine and lagoonal shales and siltstones with sandstone and tuffite interbeds. Shales and siltstones are brown to dark gray, sandy, and highly calcareous. Tuffites are crystallovitroclastic, chloritized, and partly carbonatized. Sandstones are red–brown, polymictic, fine- to medium-grained, micaceous, and locally carbonated.

Based on single finds of conchostracans *Pseudoestheria timanensis* Mol., *Loxomegaglipta* cf. *tschalyschevi* Mol., and *Leptestheria* aff. *abescunica* Mol., the Fetisovka Formation was referred for a long time to the Induan Stage (Aliev *et al.*, 1977). Our data on fossil remains from these deposits allows them to be dated reliably as late Olenekian. Ammonoids of the *Hellenites* genus (determinations by A.A. Shevyrev) found there are ascribed to the *Columbites* Zone (the uppermost Olenekian). Bivalves *Bakevellia kambei* Nakaz., *Myalina dalailamae* Vern., *Ornithopecten temirbabensis* Kipar., and *Bakevellia pannonica bogdoensis* Kipar. also indicate the Olenekian age of the host rocks. The conchostracan assemblage is represented by species *Lioestheria svetlakovae* Lop., *L. kokumbaica* Lop., *L. mokrinskyi* Mol., *Cycloestheria alievi* Lop., and *Limnadia parsimurunica* Lop. (determinations by A.Yu. Lopato), which coexist in the upper Olenekian deposits of Mangyshlak with ammonoids and bivalves. The myospore assemblage is dominated by *Densiosporites nejburgii* Balme that is characteristic of the Olenekian Stage. Based on the cited evidence, it can be suggested that the Fetisovka Formation continues the section penetrated by Borehole Ataburun P-1 (Fig. 3).

Thus, the upper Olenekian sedimentary complex in the southern flank of the Southern Mangyshlak depression is composed of the thick rock sequence deposited in the formerly extending Aksu–Kendyrli depression. This is evident from the transgressive occurrence of the upper Olenekian deposits on the surface of the folded Permian–Triassic strata in the southern part of the depression and on the eroded surface of the Paleozoic rocks (reflecting horizon F) in its northern part. Similar patterns are also observed in the central Mangyshlak paleodepression (the Zhetybai–Uzen' flexure). This allows us to suggest the existence of two spacious paleodepressions divided by the Southern Mangyshlak system of paleodomes during the considered period of time (Figs. 4, 5).

The unconformity in the upper part of the rock complex is related to the post-Middle Triassic erosion that is evident from the absence of rocks corresponding to the interval between reflecting horizons V_3 and V_3^* (Fig. 4). The stratigraphic range of the upper Olenekian is the least complete in the central part of the Southern Mangyshlak depression (Coquina Formation); in its marginal parts, it increases on account of both lower and upper strata of the rock complex. In the Zhetybai–Uzen' flexure, which has been perfectly investigated owing to drilling, the Tyurupin Group of the late Olenekian age comprises all the local biostratigraphic units defined in outcrops in the Mangyshlak Mountains, namely the *Dorikranites, Tirolites, Columbites*, and *Stacheites* zones. Thus, the combined analysis of the seismic and stratigraphic data allowed the relationships between upper Olenekian sequences of the Southern Mangyshlak to be more reliably established (Fig. 5).

The Middle Triassic seismic sequence is reliably traced across the entire Aksu–Kendyrli flexure. The thickness of corresponding rocks increases southward from 300 to 600 m. The Middle Triassic is represented by the Aksu–Kendyrli Formation (its complete section

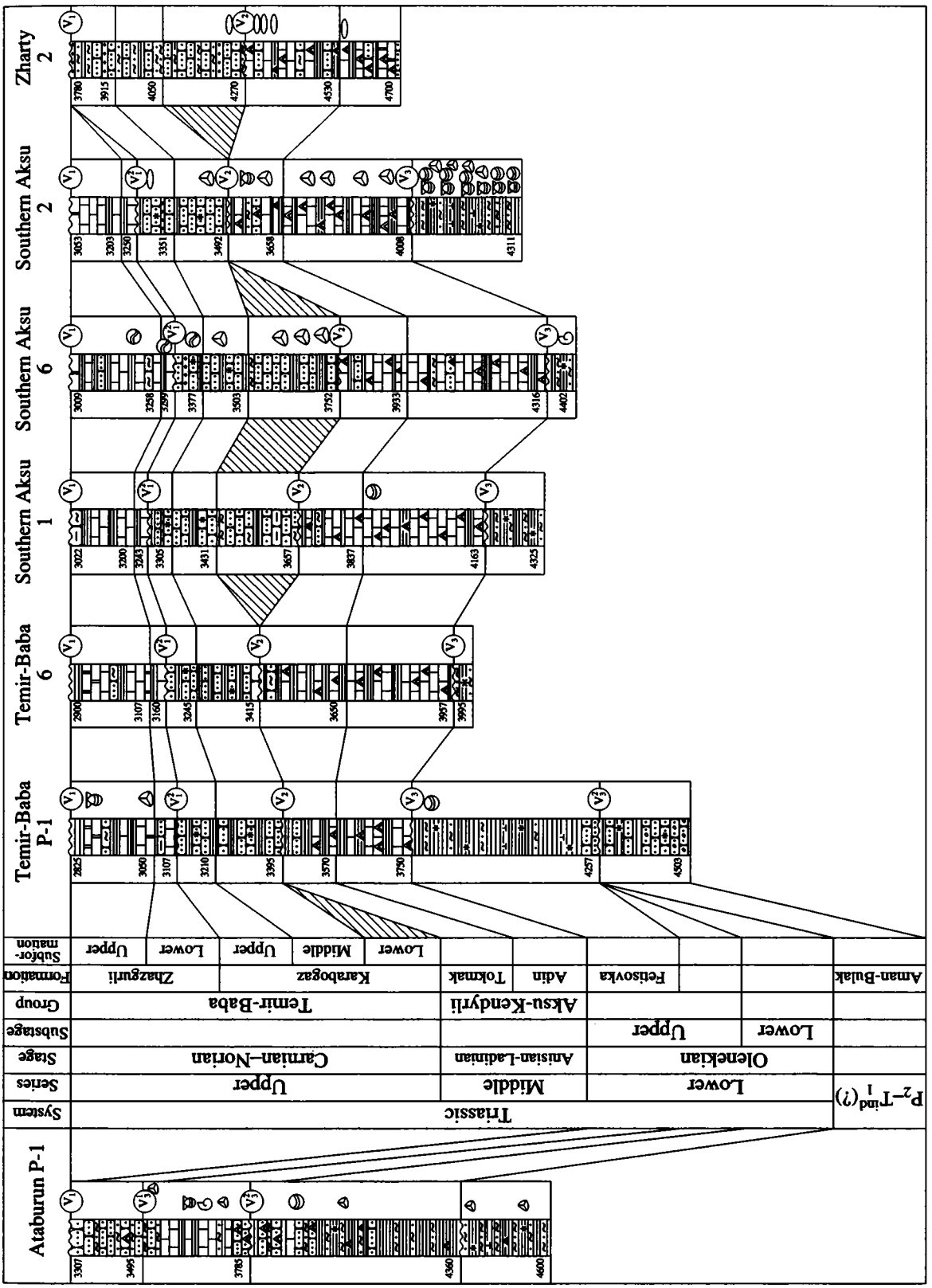

Fig. 3. Correlation of the Upper Permian–Triassic in the Aksu-Kendyrli flexure. Hatchure shows the lower formation of the Karabogaz Formation that fills the Late Triassic incision and trough in the Zhazgurli depression. For the legend see Fig. 2.

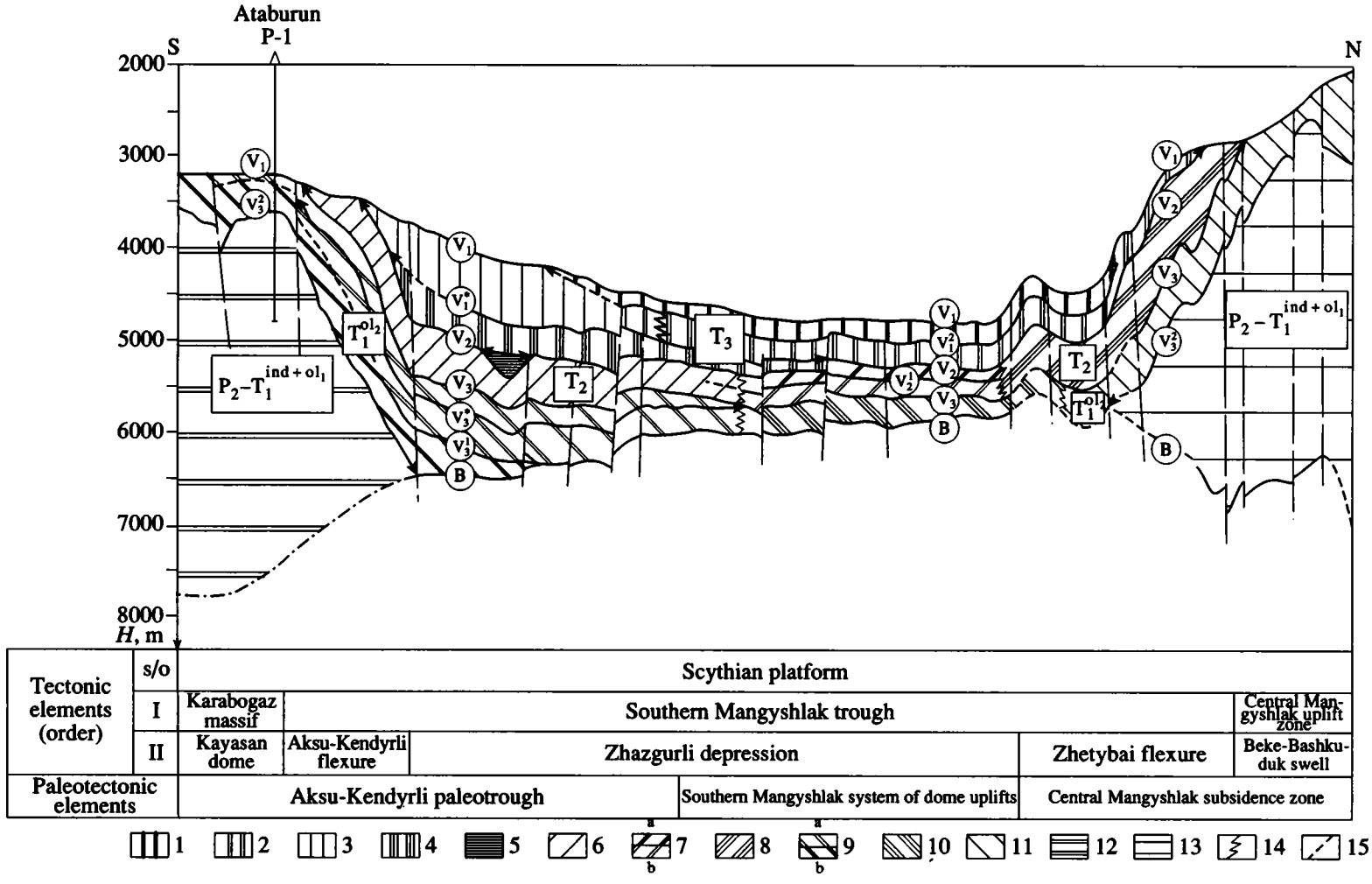

Fig. 4. Seismostratigraphic cross section along the regional profile (for location see Fig. 1). The Upper Triassic seismic sequence (continental terrigenous seismic formation): 1, mainly clayey gray-colored lacustrine and palustrine facies (Bakand Formation); 2, sandy–clayey gray-colored deposits of the alluvial plain; 3, coarse-grained red-colored facies of lateral accretion; 4, coarse-grained red-colored facies of the synsedimentary incision filling; 5, coarse-grained red-colored facies of the pre-Late Triassic incision filling; the Middle Triassic seismic sequence: 6, lagoonal clayey–carbonate gray-colored sequence (Aksu-Kendyrli Group); 7, shallow-water clayey–carbonate gray-colored sequence: a, terrigenous–tuffaceous (Tasbulat Formation); b, carbonate (Aktas and Tengin formations); 8, shelf clayey–carbonate gray-colored sequence (Southern Zhetybai Group); the upper Olenekian seismic sequence: 9, shallow-water clayey–carbonate sequence: a, mainly clayey (Fetisovka Formation); b, clayey–carbonate gray-colored facies; 10, lagoonal mainly clayey red-colored facies (Coquina Formation); 11, carbonate-clayey gray-colored facies of the open shelf (Tyururpin Group); the Upper Permian–lower Olenekian seismic sequence: 12, continental sandy–clayey red-colored dislocated sequence (Amanbulak Formation in the upper part); 13, continental sandy–clayey red-colored dislocated sequence (Dolnapin + Otpan + Birkut formations in the upper part); 14, facies transition; 15, fracture.

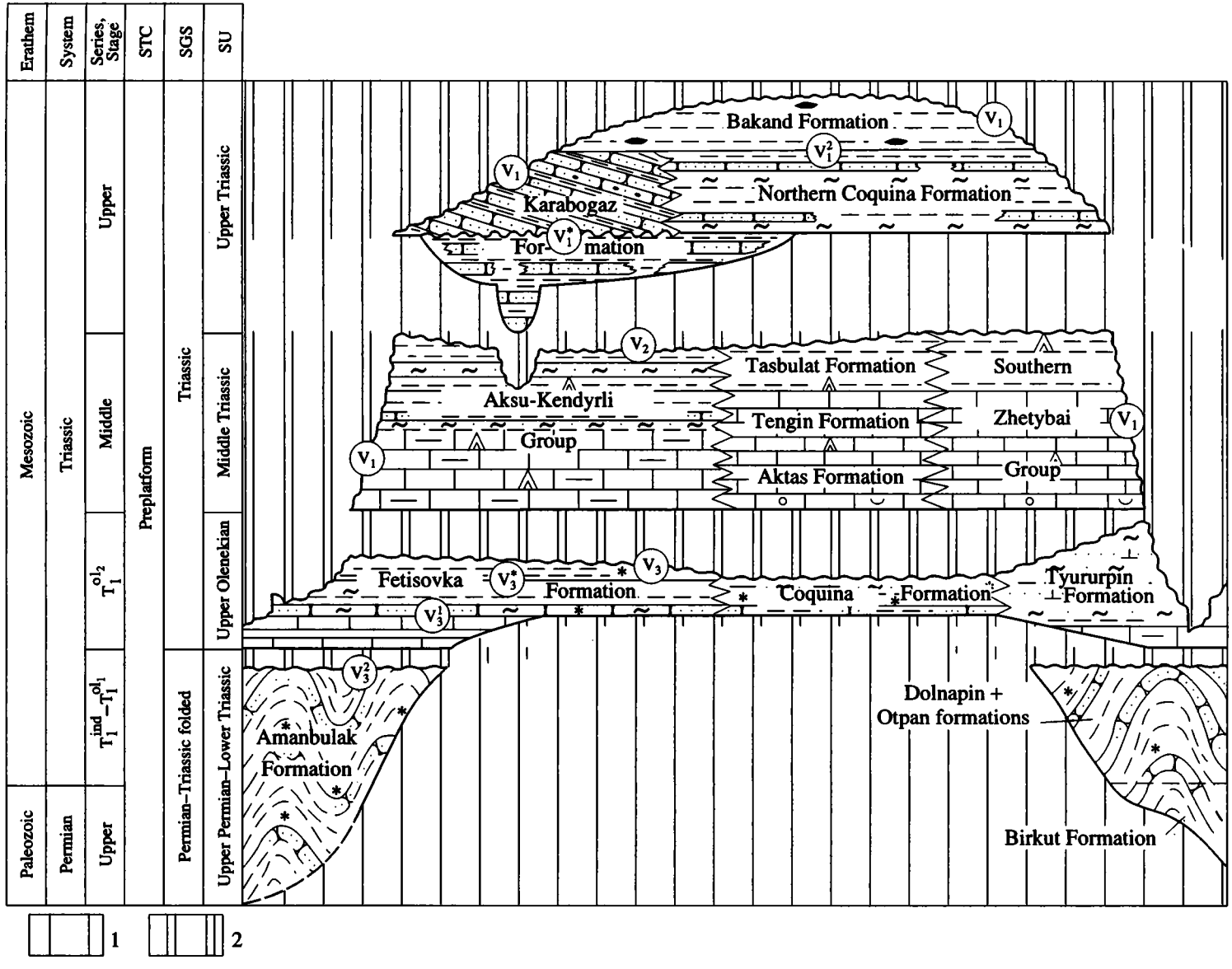

Fig. 5. Chronostratigraphic cross section along the regional seismic profile. Hiatuses related to: 1, nondeposition; 2, erosion. For other symbols see previous figures.

is penetrated by boreholes Southern Aksu-1, 2, and 6 and Temir-Baba P-1, 6, and 3; the lower part of the section was encountered by borehole Temir-Baba-5, and the upper part, by boreholes Southern Aksu-5 and Aksu-Kendyrli-14). Deposits of the considered group overlie the Fetisovka Formation and comprise the Adin and Tokmak formations. The Adin Formation is composed of gray-colored limestones with rare tuffaceous shale and siltstone interbeds. Limestones are sometimes lumpy, pelitomorphic and fine-crystalline; less commonly they are oolithic and pseudooolithic. Tuffaceous shales and siltstones are dark gray, bedded to massive, and have an admixture of pelitomorphic carbonate and inclusions of coalified plant remains and pyrite. The Tokmak Formation consists of gray-colored carbonate–tuffaceous rocks. Among its dominant rocks, there are shales and tuffaceous siltstones with fissures filled by calcite and pyrite. Clayey massive dense limestones with calcite veinlets and nests are less common. A subordinate share belongs to clayey, calcareous, and micaceous siltstones and tuffaceous siltstones with coalified plant remains, pyrite inclusions, and calcite veinlets.

The Aksu-Kendyrli Group represents a logging marker for the entire pre-Jurassic section in the considered territory (Fig. 2). It is characterized by the highest specific resistance, averaging 50 Ω/m (in the Adin Formation, it amounts to up to 70–80 Ω/m); its caliper readings are close to the nominal ones, and the unit also shows a slightly differentiated curve of spontaneous polarization (SP). In the western Aksu-Kendyrli flexure, the lower part of the Middle Triassic seismic sequence is marked by a distinct dynamically expressed three- to four-phase reflection (horizon V_2^{IV}).

The Middle (probably Early–Middle) Triassic age of the Adin Formation is established arbitrarily based on the ostracod assemblage that includes only *Renngartenella* ex gr. *aligera* Starozh., *Glorianella* sp., and *Triassinella* sp. (determinations by N.N. Starozhilova and D.A. Kukhtinov). The stratigraphic distribution of conchostracan *Lioestheria volozhi* Lop. recently found by A.Yu. Lopato in Mangyshlak is still poorly studied. The Tokmak Formation yielded bivalves *Myophoriopis gregaroides* (Phil.) characteristic of the Middle–Upper Triassic and also *Bakevellia exporrecta* (Leps.), *Unionites fassaensis* (Wissm.), and *U. canalensis* (Cat.), which are reported from the Lower-Middle Triassic. Taking into consideration these data, the formation can be dated as Middle Triassic in age.

Northward, the Aksu-Kedyrli Group is replaced by the Aktas, Tengin (both are mainly carbonate), and Tasbulat (terrigenous–tuffaceous) formations, which give way to the undivided South-Zhatybai Group farther to the north (Figs. 4, 5).

The Upper Triassic seismic sequence has the largest thickness (1200 m) in the southern flank of the Zhazgurli depression, whereas in the Aksu-Kendyrli flexure, it is observed only in the deepest areas where Upper Triassic deposits were not eroded (Figs. 1, 4). The Upper Triassic and Middle Triassic seismic sequences are separated by a distinct erosional unconformity (reflecting horizon V_2); in some areas, local incisions are observed. The seismic record of the Upper Triassic sequence reveals several seismic facies differing in reflection patterns.

The Upper Triassic deposits were drilled only in the western part of the Aksu-Kendyrli flexure, where they are represented by the Temir-Baba Group that comprises the Karabogaz and Zhazgurli formations. The Karabogaz Formation is penetrated by boreholes Temir-Baba-1, 3, and 6 and Southern Aksu-1, 2, 3, 5, and 6. It is composed of talus, lacustrine, and alluvial deposits consisting of weathering crust products and redeposited material from older Triassic strata; redeposition is evident from the presence of Early Triassic ostracods, Early and Middle Triassic myospores, and charophytes. The formation is distinctly subdivided into three subformations. The lower, mainly sandy, unit has a limited distribution and is composed of gray sandstones with interbeds of dark gray shales and siltstones. Sandstones are polymictic, silty, and irregularly calcareous and enclose coalified plant remains along bedding surfaces. The middle subformation is represented by pinkish gray arkose fine- to coarse-grained sandstones with interbeds of pinkish gray highly micaceous shales. The upper subformation consists of alternating siltstones and sandstones with shale and gritstone interbeds in its upper part. Siltstones and shales are greenish and reddish brown, bedded, micaceous, and mostly (70%) composed of weathering crust products. Sandtones are similarly colored, variably grained, and irregularly calcareous; they contain coalified plant remains along bedding surfaces and 20–50% weathering crust products.

The considerable amount of the redeposited coarse-grained and poorly sorted material suggests the dominant role of intermittent streams during the accumulation of the Karabogaz Formation deposits. A thorough correlation based on geophysical records and lithological–geochemical data revealed that the lower subformation 200–250 m thick is present only in the section of boreholes Southern Aksu-1, 6 and Zharty-2 (Fig. 3). This indicates a local distribution of talus and proluvial Upper Triassic deposits in the territory in question, although they are widespread in the subsided zone of the Zhazgurli depression, where they correspond to seismic facies of incision infilling, synsedimentary trough, and lateral accretion (Groshev, 1990). Northward, the Karabogaz Formation is replaced by alluvial and lacustrine to palustrine deposits of the Northern Coquina and Bakand formations (Fig. 5). The thickness of the Karabogaz Formation is up to 600 m. It is overlain by the Zhazgurli Formation, which is locally distributed in the western Aksu-Kendyrli flexure (boreholes Temir-Baba-1 and 6, Southern Aksu-1, 2, and 6). The formation has a two-member structure. The lower subformation is mainly represented by gray-colored

shales with sandstone and limestone interbeds; the upper one consists of intercalating gray and greenish gray limestones and dolomites with gray sandstone, siltstone, and shale interbeds. Limestones are pelitomorphic–lumpy, dolomitized, with pyrite and calcite inclusions. The thickness of the Zhazgurli Formation varies from 0 to 310 m.

The age of the Temir-Baba Formation is substantiated by its stratigraphic position and by finds of myospores *Dictiophyllum rugosum* Lindl. et Hutt. emend. Kruch., *D. vulgaris* (Mal.) Kruch., *Cyathidites crassiangulatus, Duplexisporites guratus, Styxisporites cooksonae* Kl., and *Cyclotriletes pustulatus* Madl. (determinations by G.M. Romanovskaya), which appear in the Late Triassic. Nevertheless, the Middle Triassic bivalve assemblage of *Myophoriopis gregaroides* (Phil.) and *M. nuculiformis* (Zenk.) previously reported by L.D. Kiparisova from the Zhazgurli Formation (Borehole Temir-Baba P-1) does not allow this part of the section to be dated unambiguously as the Late Triassic.

CONCLUSION

Proceeding from the aforesaid, we may conclude that, during the entire Late Permian–Triassic, specific sedimentation environments were responsible for the accumulation of peculiar sedimentary formations in the Aksu-Kendyrli flexure area. At the end of the Early Permian, the crustal consolidation terminated in Mangyshlak, and this area entered the period of continental development (Volozh, 1991). In the pre-late Olenekian, the shallow-water areas of the Aksu-Kendyrli intracontinental basin, which experienced subsidence at that time, accumulated thick sandy–clayey sediments (the Aman-Bulak Formation) under conditions of an arid climate and intense influx of clastic material from the nearby provenance (the Karabogaz Paleozoic massif). The late Olenekian time was marked by the ingression of the Tethyan waters into the Mangyshlak region, and the Aksu-Kendyrli area was occupied by a shallow-water basin, as is evident from the ammonoid and bivalve fossils. First carbonate and later clayey–silty deposits (the Fetisovka Formation) were formed in this basin. A similar trend of sedimentation against the background of a general deepening of the basin was retained during the Middle Triassic, when thick gray-colored clayey–carbonate deposits (the Aksu-Kendyrli Group) were accumulated there. The Late Triassic epoch in the larger part of the Mangyshlak area was marked by the formation of terrigenous sequences in continental environments and under conditions of progressing humidization of the climate. At the beginning of this epoch, the deeper parts (erosional incisions) of the Aksu-Kendyrli paleodepression were filled with coarse sediments, which were later overlain by more widespread Late Triassic deposits of the synsedimentary filling type. Later, red-colored sequences of the lateral accretion were deposited. Their formation occurred in the shallow-water continental basins with high-energy hydrodynamics and an intense supply of erosion products from the Paleozoic Karabogaz massif (the Karabogaz Formation). During the final stage of the Late Triassic deposition, waters of the paleo-Caspian Sea probably entered the Aksu-Kendyrli flexure area again. This event resulted in the formation of carbonate–terrigenous deposits (the Zhazgurli Formation).

Reviewers A.S. Alekseev and A.A. Shevyrev

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Melekhovo Horizon: A New Stratigraphic Unit of the Gzhelian Stage, Upper Carboniferous (East European Platform)

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Abstract—The decision of March 9, 1992, of the Bureau of the Interdepartmental Stratigraphic Committee to change the position of the Carboniferous–Permian boundary, and, hence, of the upper Gzhelian boundary, in the general stratigraphic scale necessitated the identification of a new Gzhelian horizon in a stratotype locality of the Russian Platform. A stratotype was chosen in the Melekhovo–Fedotovo quarry, in outcrops from the Moshachikha ravine, and in an open pit near the Klyazminskii Gorodok village, all in the vicinity of Kovrov City, Vladimir District, the Oka–Tsna arch. The new unit is suggested to be termed as the Melekhovo Horizon. Type sections with fusulinids and conodonts were described bed by bed in outcrops and boreholes. The boundary deposits between the Noginsk and newly identified Melekhovo horizons were correlated in the Moscow, Ivanovo, and Yaroslavl districts and in Tatarstan. New fusulinid species are described, and photomicrographs of typical species from the *Daixina bosbytauensis–D. robusta* Zone are presented.

INTRODUCTION

The decision of March 9, 1992, of the Bureau of the Interdepartmental Stratigraphic Committee (ISC) (*Postanovleniya ...*, 1992) to change the position of the boundary between the Carboniferous and Permian systems in the general stratigraphic scale demanded that a new horizon be identified in the Gzhelian Stage. According to this decision, a boundary between these systems, and, hence, the base of the Asselian Stage, was placed in the goniotite scale between the *Shumardites–Vidrioceras* and *Svetlanoceras juresanites* genozones in accordance with a species change within a phylogenetic *Artinskia irinae–A. kazahstanica* lineage. In the fusulinid scale, this level corresponds to a boundary between the *Daixina bosbytauensis–D. robusta* and *Sphaeroschwagerina vulgaris–S. fusiformis* (s. str.) zones. Its relation with the conodont scale was not fully ascertained. A boundary between the Carboniferous and Permian systems was placed between the *Streptognathodus wabaunsensis* and *S. cristellaris* zones. This boundary stratotype was established in the Aidaralash section (South Urals) between Bed 19/5 and Bed 20 (*Postanovleniya ...*, 1985).

The range of the Gzhelian Stage was delimited by S.N. Nikitin as far back as 1890. He described its main outcrops (stratotypes) in the areas of the towns Gzhel and Pavlovskii Posad, Rusavkino and Amerevo villages, and the town of Noginsk, formerly Bogorodsk (Nikitin, 1890). The names of these localities were later used by Dan'shin (1947) to name the sequences he identified in the Gzhelian Stage as the Rusavkino, Shchelkovo, Amerevo, Pavlovskii Posad, and Noginsk sequences. Only a few brachiopod species, indicative of the Late Carboniferous age (Ivanova and Rozovskaya, 1967), were known from the upper Noginsk Sequence, which was correlated with the Orenburgian Stage or with the *Daixina sokensis* Zone in the stratigraphic scale for the Russian Platform unified in 1962 (*Resheniya ...*, 1965). By 1975, geologists of the Central Russia Geological Survey discovered numerous zonal *Daixina sokensis* forms and a fusulinid assemblage of the *Daixina sokensis* Zone in core samples from the Noginsk Sequence penetrated by a few boreholes drilled in the Noginsk area. Because this stratotype ascribed the sequence to the Gzhelian Stage, no grounds remained to refer it to the Orenburgian Stage. These data were sufficient to rank the Noginsk Sequence as a horizon completing the Gzhelian Sequence in the type locality (Makhlina *et al.*, 1975). Because the outcrops of the Pavlovskii Posad and Noginsk deposits had not survived, the sequence of Hole 7k drilled in the Noginsk area was offered as a type section (neostratotype) of the Noginsk Horizon.

NEW HORIZON OF THE GZHELIAN STAGE

According to the stratigraphic scale for the Carboniferous of the Russian Platform that was unified in 1988 (*Reshenie ...*, 1990), the Gzhelian Stage comprised three horizons: the Dobryatinskii Horizon (comprising the Rusavkino, Shchelkovo, and Amerevo beds) and the Pavlovskii Posad and Noginsk Horizons. The overlying beds corresponding with the *Daixina bosbytauensis* Zone were classified as the lower member of the

Kholodnyi Log Horizon (*Schawagerina vulgaris–Sch. fusiformis* s. l. Zone) of the Asselian Stage. This segment of the section was penetrated as a part of the Noginsk Horizon stratotype by Hole 7k, which was visited during field trips by the participants of the 8th International Carboniferous Congress (*Putevoditel'...*, 1975). The same sequence is exposed on the Oka–Tsna arch in the Melekhovo–Fedotovo quarry not far from Kovrov City, Vladimir District, where we studied it in detail (Makhlina *et al.*, 1975, 1979; Makhlina and Isakova, 1984). However, as was mentioned above, the *Daixina bosbytauensis–D. robusta* Zone was included in the Gzhelian Stage by the ISC decision of 1992. This necessitated the introduction of a new horizon corresponding with the Gzhelian *Daixina bosbytauensis–D. robusta* Zone in the stratotype locality of this stage. The upper contact of the Noginsk Horizon, or the lower boundary of the newly identified horizon, bordering the Asselian Stage (*Sphaeroschwagerina vulgaris–S. fusiformis* Zone), is fairly distinct in the sequence penetrated by Hole 7k and also in the section exposed in the Melekhovo–Fedotovo quarry. In this quarry, the Noginsk Horizon is represented by interbedded massive and thin-flagged, micro- to fine-grained dolomites and light gray dolomitized foraminiferal limestones containing fusulinids of the *Daixina sokensis* Zone. The fusulinid assemblage includes the following (Samples 52, 7–10 in Fig. 2): *Daixina sokensis* (Raus.), *D. fragilis* Ros., *Jigulites jigulensis* (Raus.), *J. volgensis* (Raus.), *J. intermedius* (Ros.), *J. major* (Ros.), *J. dagmarae* Ros., *Triticites paraarcticus* Raus., *T. condensus* Ros., *T. morkvashensis* Ros., *Rugosofusulina praevia orientale* Bensh, *Quasifusulina longissima* (Moeller), and *Q. eleganta* (Shlyk.). Species of the genus *Adetognathus* and also *Streptognathodus elongatus* Gunn. were identified in the Noginsk rocks from Hole 7k (Samples 7k-11-09 and 7k-9-010, depth levels of 59.3–61.3 m). A species intermediate between *Streptognathodus elongatus* and *St. wabaunsensis* (Sample 7k-13-09, depth ?) was found at the top of the horizon. Conodonts were identified by A.S. Alekseev and A.N. Reimers (Collection 4612a stored at the Department of Paleontology, Moscow State University).

In addition to these, O.L. Kossova collected corals from the lower beds of the Noginsk Horizon and identified them as *Gzhelia rouilleri* Stuck. and *Arctophyllum* aff. *intermedium* (Toula). The sequence of this horizon is completed by a distinct marker bed (Bed 27, 0.8 m thick) of poorly colored, unfossiliferous clayey dolomite or variegated dolomitic marl (blue marl). Below this marl, Kossova found corals (Sample 12) and identified them as *Gzhelia rouilleri* Stuck., *Arctophyllum minima* Koss., and *Lophocarinophyllum* sp. She found that the stratigraphic interval of *Gzhelia rouilleri* Stuck. ranged from the base of the Gzhelian Stage to its contact with the Lower Permian. Earlier she had believed that *Arctophyllum minima* Koss. appeared at the base of the *Daixina bosbytuaensis–D. robusta* Zone (Davydov and Dorofeeva, 1991, Fig. 1).

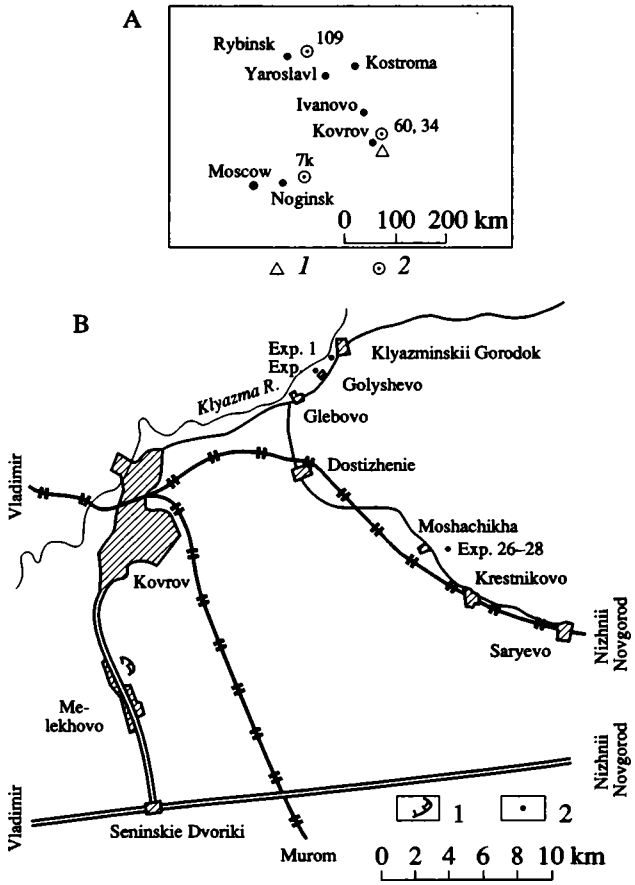


Fig. 1. Location of studied sections of the Melekhovo Horizon. A, stratotype localities (*1*) and boreholes (*2*); B, quarries (1) and outcrops (2).

COMPOSITE STRATOTYPE OF THE MELEKHOVO HORIZON

In our opinion, the deposits that lie above the blue marl should be ranked as a new horizon of the Gzhelian Stage. We proposed to name it Melekhovo Horizon after the Melekhovo settlement, near which the Melekhovo–Fedotovo quarry is located. We suggested that a sequence embracing sections from the Melekhovo–Fedotovo quarry (outcrops in the Moshachikha ravine) and an open pit near the Klyazminskii Gorodok settlement could be recognized as a holostratotype of the new horizon (Figs. 1, 2). This was discussed and approved on October 26, 1994, by the Regional Interdepartmental Stratigraphic Committee for the central and southern Russian Platform. In the Melekhovo–Fedotovo quarry, the basal bed of the horizon is an organic limestone (Bed 28, 0.5 m thick), highly dolomitized, slightly clayey, with a bluish to pinkish tint. The limestone is a fine-grained rock containing echinoid remains and solitary corals. The basal surface of this bed is uneven. Upward follows a layer (Bed 29, 1.5 m thick) of white, massive, fine-grained dolomite, which is superseded by a layer (Bed 30, 1.0 m thick) of

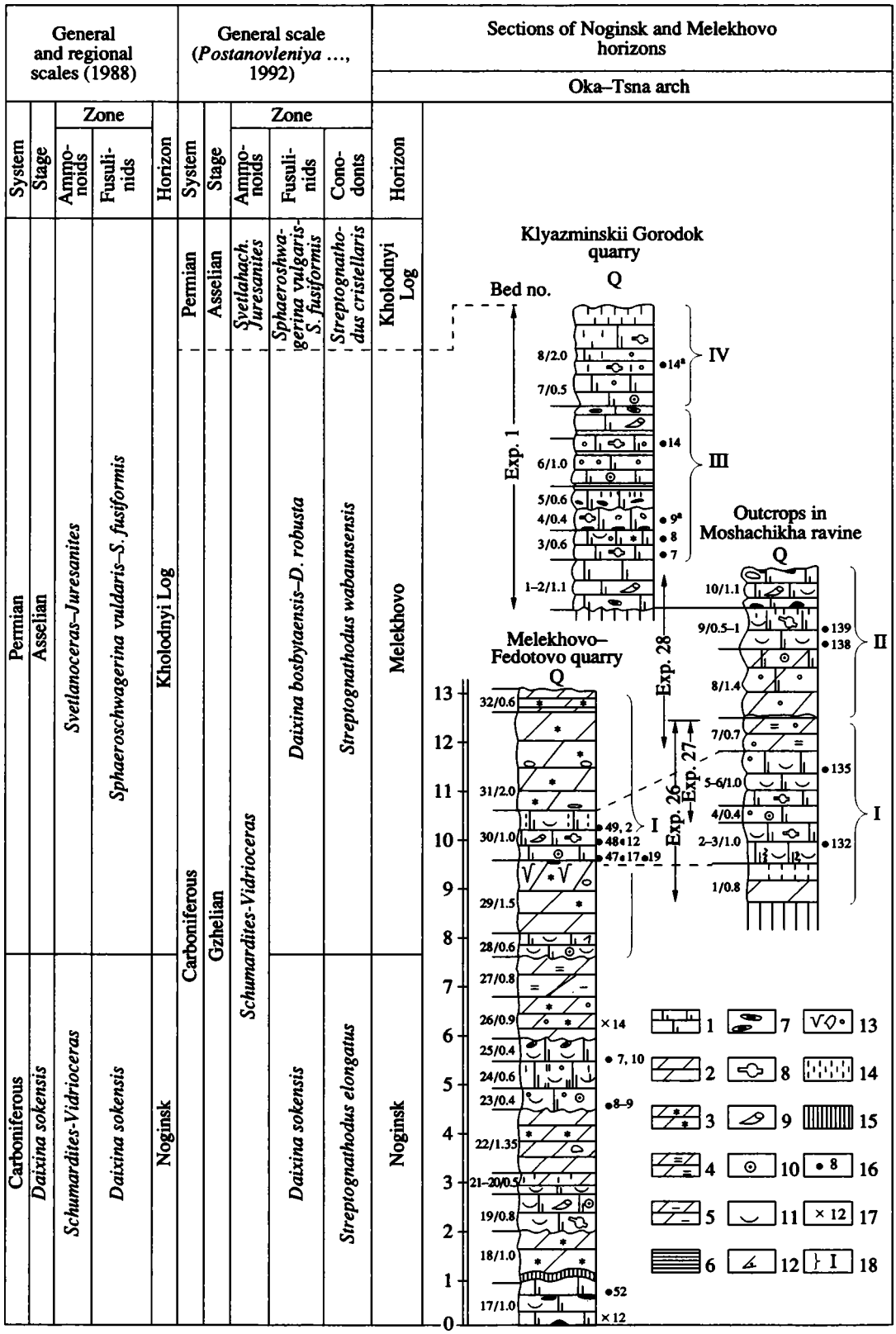


Fig. 2. Composite stratotype of the Melekhovo Horizon, Gzhelian Stage: (1) highly dolomitized limestone (secondary dolomite); (2) dolomite; (3) fine-grained dolomite (limestone); (4) clayey dolomite; (5) dolomitic marl; (6) clay; (7) cherty nodules and silicified rocks; (8) fusulinids; (9) corals; (10) crinoids; (11) detrital limestone; (12) calcite and calcite geodes; (13) vertical tubes (cavities), leaching caverns and voids; (14) dolomite meal; (15) rock waste; (16) sample containing fusulinids; (17) sample containing corals; (18) rhythm interval.

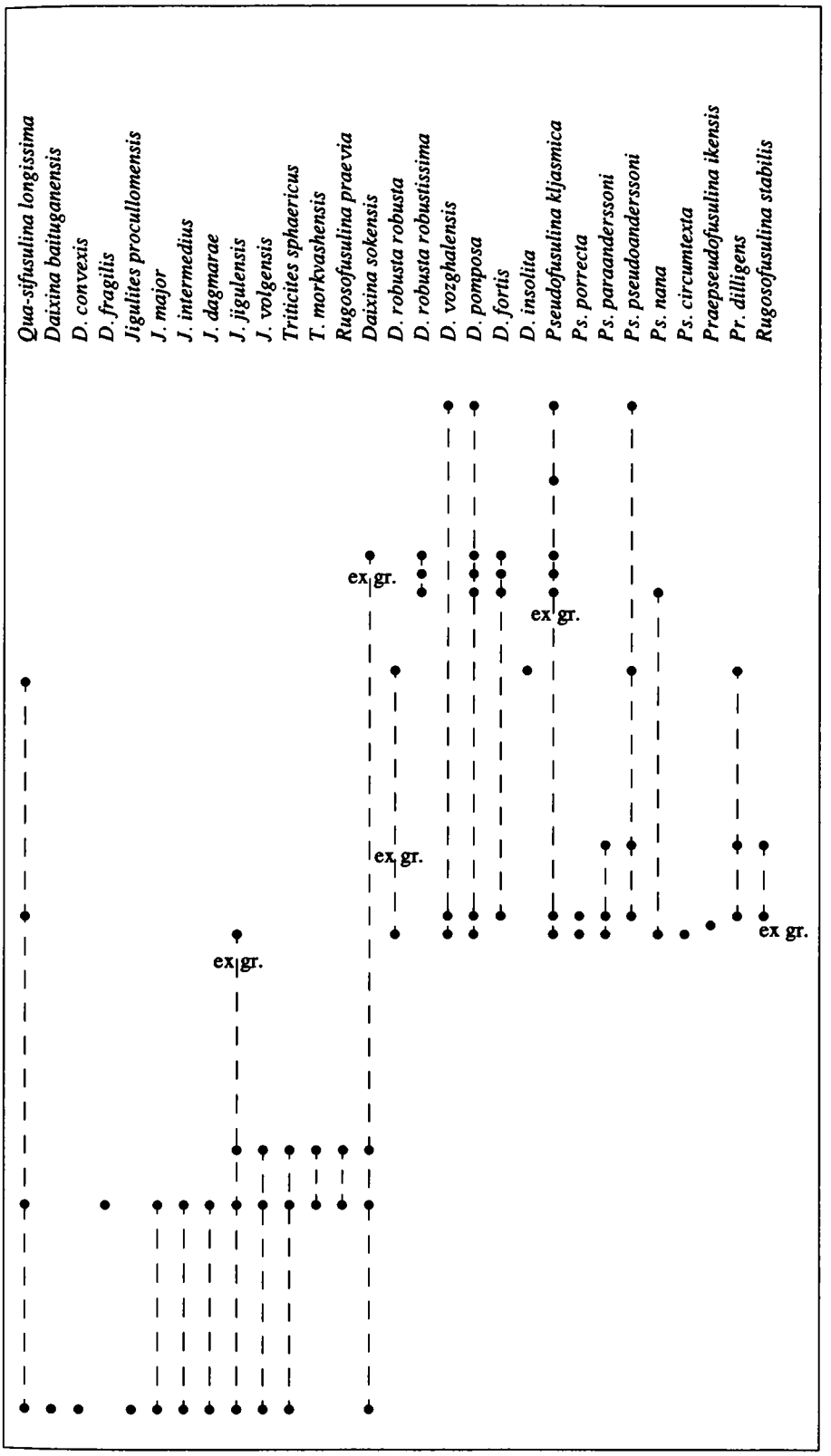
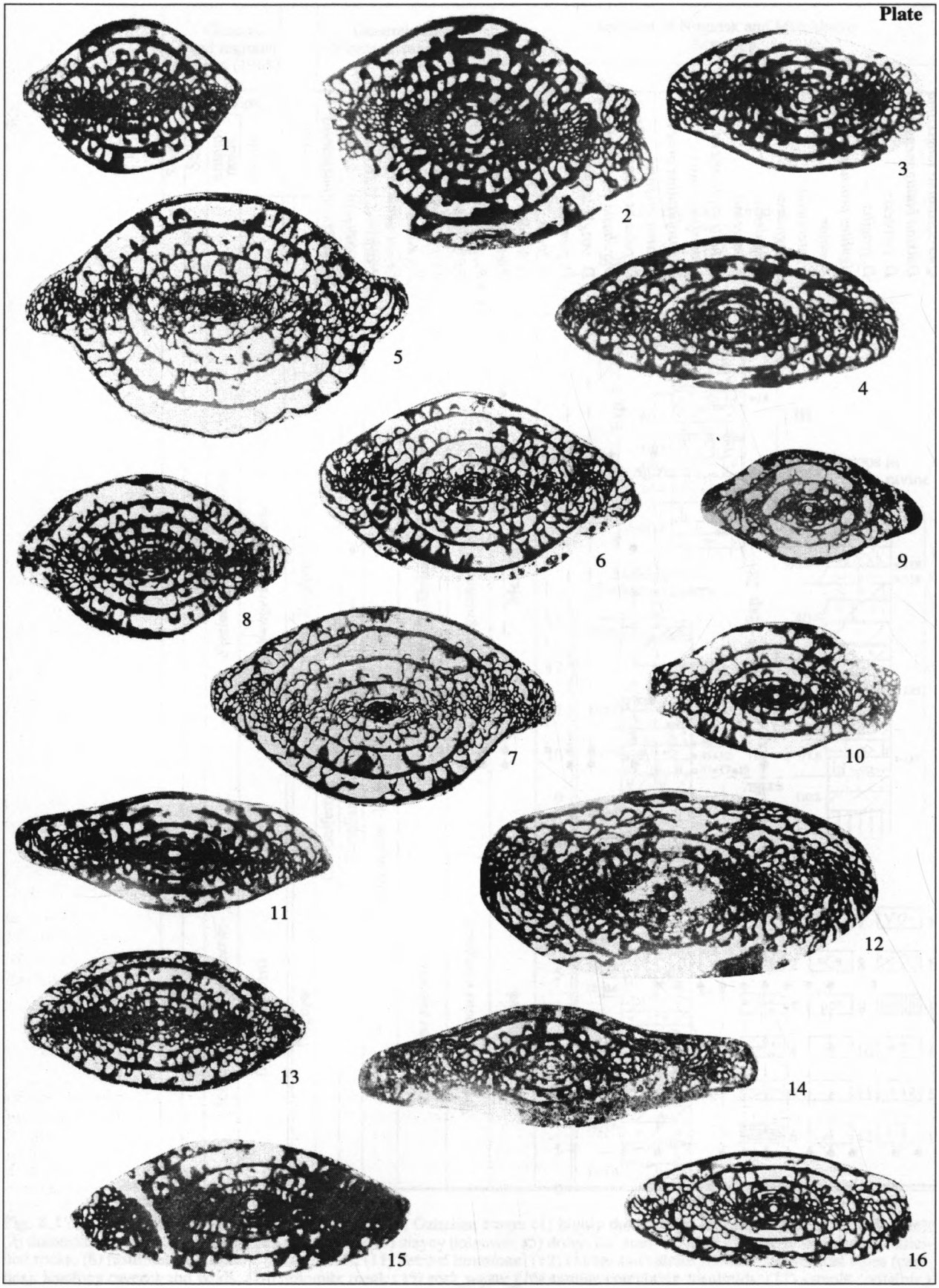

Qua-sifusulina longissima
Daixina baituganensis
D. convexis
D. fragilis
Jigulites procullomensis
J. major
J. intermedius
J. dagmarae
J. jigulensis
J. volgensis
Triticites sphaericus
T. morkvashensis
Rugosofusulina praevia
Daixina sokensis
D. robusta robusta
D. robusta robustissima
D. vozghalensis
D. pomposa
D. fortis
D. insolita
Pseudofusulina kljasmica
Ps. porrecta
Ps. paraanderssoni
Ps. pseudoanderssoni
Ps. nana
Ps. circumtexta
Praepseudofusulina ikensis
Pr. dilligens
Rugosofusulina stabilis
ex gr.
ex gr.
ex gr.
ex gr.
ex gr.

Fig. 2. (Contd.)



Plate
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secondary dolomite or highly dolomitized limestone with solitary corals and diplopores, and by intercalations of crinoidal–algal limestone containing fusulinids characteristic of the *Daixina bosbytauensis–D. robusta* Zone (Samples 47–49, 2, 12, 17, 19), such as *Daixina robusta* Raus., *D. vozghalensis* Raus., *D. pomposa* Sjom., *D. fortis* Sjom., *Pseudufusulina paraanderssoni* Raus., *Ps. nana* Sjom., *Ps. kljasmica* (Sjom.), *Ps. circumtexta* Scherb. (Plate 1, Fig. 12), *Ps. porrecta* (Sjom.) (Plate 1, Fig. 14), *Praepseudofusulina ikensis* Dobr. (Plate 1, Fig. 11), and *Triticites* ex gr. *plummery*. The secondary dolomite is overlain by massive and thin-flagged dolomites (Bed 31, 2 m thick, and Bed 32 with an incomplete thickness of 0.5 m, respectively). The thickness of the Melekhovo Horizon exposed in the quarry is 5.6 m. Its top was obliterated by Quaternary erosion.

A considerable portion of the composite holostratotype of the Melekhovo Horizon is exposed in outcrops and quarries located within the Oka–Tsna arch: along the western side of the Moshachikha ravine near the Moshachilkha village, i.e., 200 m away from the point where the ravine is intersected by a highway, and in an abandoned stone quarry near the Klyazminskii Gorodok village, 50 m from a weaving factory. These outcrops were studied by Semina (1961), who compiled a composite section. In 1973, this section was studied by Makhlina and Kulikova (Makhlina *et al.*, 1975), and in 1974–1976, by T.N. Isakova. The lower half of the Melekhovo Horizon is exposed in three outcrops in the Moshachikha ravine (Sites 26–28 in Fig. 1). Beds 2–6 (2.4 m thick) at Sites 26 and 27 of the Moshachikha ravine correlate with Bed 30 of the Melekhovo–Fedotovo quarry. The beds consist of secondary dolomites or highly dolomitized organogenic limestones containing fusulinids, solitary corals, echinoid fragments, and other organic remains. The rocks are yellow to light yellow and thin-flagged, with voids produced by the leaching of organic detritus and fusulinids. The rocks are partially silicified in the lower part of the section, where they are leached and calcitized almost ubiquitously.

According to our data, this part of the section bears the following fusulinids (Samples 132 and 135): *Daixina* ex gr. *robusta* Raus., *Pseudofusulina pseudoanderssoni* Sjom., *Praepseudofusulina dilligens* Isak. sp. nov. (Plate 1, Fig. 19), *Ps. paraanderssoni* Raus. (Plate 1, Fig. 13), and *Rugosofusulina* ex gr. *stabilis* (Raus.). According to Semina (1961), the assemblage also includes *Quasifusulina longissima* (Moell.), *Rugosofusulina* ex gr. *serrata* Raus., *Pseudofusulina gregaria shustovensis* Raus., *Ps. krotowi globulus* Raus., *Schwagerina* sp., and also small foraminifers represented by numerous *Textularia* and *Tetrataxis*. The latter group is more common for the deposits of the Asselian Stage. It is not unlikely that, because of the poor preservation of the fusulinid fauna in this locality, Semina mistook some species of the *Daixina robusta* group for *Globofusulina globulus* (Raus.) (*Pseudofusulina krotowi globulus* after Semina), which have subspherical shells typical of the former. Bed 7 (0.7 m thick) consists of yellow, fine-grained, slightly clayey, poorly cemented dolomite with a small amount of quartz grains. Its top interval includes an interlayer of brown calcareous clay with quartz geodes.

Beds 8–10 are composed of white secondary dolomite or dolomitized limestone containing interlayers of fusulinid limestone, often with tubular bryozoans. The rock is cavernous because of numerous voidal detritus and fusulinids. The top of Bed 9 (0.5–1 m thick) includes a layer of dolomite meal; the bottom of Bed 10 contains gray chert nodules. Bed 10 terminates the sequence, has an apparent thickness of 1.1 m, and consists of brownish yellow, laminated, very fine- to fine-grained dolomite. This part of the sequence (Samples 138 and 139) contains the following fauna assemblage: *Praepseudofusulina dilligens* Isak. sp. nov. (Plate 1, Fig. 9), *Pseudofusulina pseudoanderssoni* Sjom., *Ps. pseudoanderssoni* aff. *latiterminosa* Sjom., *Ps. paraanderssoni* Raus., *Daixina* ex gr. *vozghalensis* Raus., *D. robusta* Raus. (Plate 1, Fig. 1), *D. insolita* Isak. sp. nov. (Plate 1, Fig. 9), and *Rugosofusulina* sp. A fragment of several coils of a poorly preserved *Schwagerina* shell was found in Sample 139. The same thin section contains *Daixina insolita* Isak. sp. nov. This assemblage can be supplemented with the following species that were identified by Semina: *Quasifusulina longissima* (Moell.), *Q. tenuissima* (Schellw.), *Triticites plummeri* Dunb. et Condra, *Pseudofusulina rostovzevae* Sjom., *Ps. paraanderssoni* Raus., *Ps. bina* Sjom., and *Ps. nana* Sjom.

Plate 1. Fusulinids of the Melekhovo Horizon (magnification ×10): (1) *Daixina robusta* Raus., Specimen no. 4612/1, Kovrov area, head of Moshachikha ravine, Sample 138; (2) *Daixina pomposa* Sjom., Specimen no. 4612/2, Kovrov area, Klyazminskii Gorodok village, Sample 7; (3) *Daixina vozghalensis* Raus, Specimen no. 4612/3, Kirzhach area, Hole 7k, Sample 7/43; (4) *Daixina insignis* Jagof., Specimen no. 4612/4, Kirzhach area, Hole 7k, Sample 7/43; (5–8) *Daixina insolita* Isakova sp. nov.: (5) holotype no. 4612/5, Kovrov area, right bank of the Klyazma River, Golyshevo village; (6, 7) type species, nos. 4612/6 and 4512/7, the same locality (possibly *Schwagerina vulgaris–Sch. fusiformis* Zone); (8) type species, Specimen no. 4612/8, Kovrov area, head of Moshachikha ravine, Sample 138; (9, 10) *Praepseudofusulina diligens* Isakova, sp. nov.: (9) holotype no. 4612/9, Kovrov area, head of the Moshachikha ravine, Sample 139; (10) type species no. 4612/10, same locality, Sample 132; (11) *Praepseudofusulina ikensis* Dobr., Specimen no. 4612/11, Kovrov area, Melekhovo Quarry, Sample 19; (12) *Pseudofusulina circumtexta* Scherb., Specimen no. 4612/12, Kovrov area, Melekhovo Quarry, Sample 7; (13) *Pseudofusulina pseudoanderssoni* Sjom., Specimen no. 4612/13, Kovrov area, head of Moshachikha ravine, Sample 138; (14) *Pseudofusulina porrecta* Sjom., Specimen no. 4612/14, Kovrov area, Melekhovo Quarry, Sample 2; (15) *Pseudofusulina paraanderssoni* Raus., Specimen no. 4612/15, Kovrov area, head of Moshachikha ravine, Sample 132; (16) *Pseudofusulina kljasmica* (Sjom.), Specimen no. 4612/16, Kirzhach area, Hole 7k, Sample 7/12.

Proceeding from the above description, two elementary rhythms, Beds 1–7 and Beds 8–10, composed mainly of organogenic, highly dolomitized limestone, can be recognized in the lower half of the Melekhovo Horizon; each of the rhythms is terminated by laminated dolomite or claystone (Beds 7 and 10).

The upper part of the Melekhovo Horizon is exposed in an abandoned quarry at the eastern side of the Klyazma River near the Klyazminskii Gorodok settlement (Beds 1–8 in Fig. 2), where Beds 1–5 (2.7 m thick) consist of yellow dolomitic limestone including 3 or 4 layers of dark chert. The limestone (Beds 3 and 4) is a cavernous rock containing fusulinids, solitary corals, and gastropod casts. In Samples 7, 8, and 9a, we determined *Daixina robusta robustissima* Sjom., *D. composa* Sjom. (Plate 1, Fig. 2), *D. fortis* Sjom., *Pseudofusulina nana* Sjom., *Ps. polymorpha* Sjom., and *Ps.* ex gr. *kljasmica* Sjom. The following species from Semina's determinations can be added to this list: *Pseudofusulina* ex gr. *krotowi* (Schellv.), *Ps.* ex gr. *gregaria* Lee, *Ps.* ex gr. *regularis* (Lee non Schellw.), *Ps.* aff. *differta* Sham., *Ps.* ex gr. *rhomboides* Sham. et Scherb., *Ps.* aff. *declinata* Sham., *Tetrataxis* sp., *Bradyina* sp., and *Textularia* sp. Upward (Beds 6 and 7, 1.5 m thick, there is light gray, fine-grained dolomitized limestone containing fusulinids solitary corals, and light gray claystone intercalations. The rock has caverns after leached fusulinids and is silicified in the top of Bed 7. The newly found fusulinids (Sample 14) are *Pseudofusulina* ex gr. *kljasmica* Sjom., *Ps. pseudoanderssoni latiterminosa* Sjom., and *Ps.* cf. *rostovzevae* Sjom.

The sequence is crowned with Bed 8 (apparent thickness 2 m) of yellow, highly dolomitized, fusulinid-bearing limestone. The rock is thin-flagged and severely fractured; some of its layers were replaced by dolomite meal. The fusulinid assemblage includes (Sample 14a) *Daixina* ex gr. *vozghalensis* Raus., *D. pomposa* Sjom., *Pseudofusulina pseudoanderssoni* Sjom., *Ps. pseudoanderssoni latiterminosa* Sjom., and *Ps. exuberata* Sham et Scherb. In addition, Semina (1961) reported *Daixina fortis* Sjom., *D. acris* Sjom., *Pseudofusulina kljasmica* (Sjom.), *Ps. regularis* (Lee non Schellw.), *Ps.* aff. *pseudopointeli* Raus, *Ps.* ex gr. *gregarea* Lee, *Pseudoschwagerina* sp., and small foraminifers from the top of the same section.

The upper boundary of the Melekhovo Horizon was not established in the section of the Klyazminskii Gorodok locality. It can be located tentatively at the base of a small pile of rock fragments near the Golyshevo village, where the following fusulinids of Asselian age were found (Sample 15): *Daixina* aff. *robusta* Raus., *D. insolita* Isak. sp. nov., *D.* cf. *pomposa* Sjom., *Schwagerina shamovi* Scherb., and *Quasifusulina longissima* (Moell.).

Similar to the lower half of the horizon, its upper half appears to consist of two rhythms as well (Beds 1–7 and Bed 8). The bottom of the lower rhythm is limestone with chert intercalations; the top is marked by limestone with clay interlayers. The upper rhythm is represented by its basal bed alone (other units seem to be missing) and consists of organogenic limestone.

Based on the evidence used to compile a composite holostratotype, the thickness of the Melekhovo Horizon is approximately 13 m in this locality.

Generally, our data agree with Semina's determinations; in some cases they supplement and add details to the fauna assemblages and the successions of the rocks.

REGIONAL CORRELATION OF GEOLOGIC SECTIONS WITH THE MELEKHOVO HORIZON STRATOTYPE

The lower and, in some areas, the upper boundary of the Melekhovo Horizon were traced in a number of boreholes (7k, 34, 60, 109) that were drilled in the Moscow, Ivanovo, and Yaroslavl districts (Makhlina *et al.*, 1975). The Melekhovo–Fedotovo, Moshachikha, and Klyazminskii Gorodok sections of the Oka–Tsna arch were correlated with the following intervals of the holes: 44.0–59.3 m in Hole 7k, 132.0–150.0 m in Hole 34, 144.0–158.0 m in Hole 60, and 204.0–217.0 m in Hole 109.

The rocks of the Noginsk and Melekhovo horizons of the Gzhelian Stage are represented in these holes by facies that are identical to those of the surface rocks in their type localities. These are secondary dolomites and highly dolomitized limestones interlayered with fine-grained dolomites and containing fusulinids and crinoid–brachiopod detritus. Fusulinids of the *Daixina sokensis* and *D. bosbytauensis–D. robusta* zones were determined in their core samples. The fossils were replaced by gypsum, as was a considerable volume of the rock itself. The core samples recovered in Hole 7k from the base of the Melekhovo Horizon (Sample 7k-7-2, interval 59, 3–61 m) contain conodonts: *Streptognathodus simplex* Gunn., *St. elongatus* Gunn., *St. nodulinearis* Cher. et Resh., *St. wabaunsensis* Gunn., and *St. alius* Achmet. The cores from the rocks above this horizon (Samples 7k-2-013 and 7k-1-014, int. 56, 5–59, and 3 m) contain *Streptognathodus simplex* Gunn., *St. wabaunsensis* Gunn., and *St.* ex gr. *wabaunsensis* Gunn. It is significant that the boundary beds between the Melekhovo Horizon and Asselian Stage do not contain fusulinids of the *Sphaeroschwagerina vulgaris* and *S. fusiformis* zones. It is not unlikely that the deposits of the lower Asselian zone, corresponding with the bottom of the Kholodnyi Log Horizon, are missing there.

The Melekhovo Horizon can be traced as far as the Volga–Ural region. Hole 5, drilled in the southeast of the Tatarstan Republic, near the Shegurcha village of the Novosheshminskii district, penetrated the Noginsk–Melekhovo boundary rocks containing fusulinids of the *Daixina sokensis* and D. *bosbytauensis–D. robusta* zones. The zonal assemblage of the former was determined at a depth of 316–317 m. It consists of numerous

species of the *Daixina sokensis* group: *Daixina sokensis* index species, *D. recava* Zolot., *D. krushiensis bella* Echl., *D.* aff. *admirabilis* Echl., and *D. dualis* Echl. The assemblage also includes various species of the *Jigulites* group: *Jigulites jigulensis* (Raus.), *J. volgensis* (Raus), *J. volgensis sizranica* Shlyk., *J. longus* Ros., and *J. intermedius* Ros. There are also species of the *Rugosofusulina* group: *Rugosofusulina stabilis* (Raus.), *R. stabilis longa* (Raus.), *R. uralensis* Ros., and *R. valida volgensis* Isak.

Along with the fusulinid assemblage of the *Daixina sokensis* Zone, this interval of the section (depth 320–322 m) contains conodonts: *Streptognathodus* Gunn., *S. elegantulus* Stauf. et Plum., *S.* aff. *wabaunsensis* Gunn., *S. alius* Achmet., and *Gondolella* sp. (collected and determined by A.N. Reimers).

A significant change in the fusulinid assemblage was observed at a depth of 307 m, where species of the *Daixina bosbytauensis* group with a broad free spiral appear and grow rapidly in abundance. The rocks penetrated in a depth interval of 301–307 m yielded a typical zonal assemblage of the *Daixina bosbytauensis–D. robusta* fusulinid zone. They contain *Daixina* aff. *bosbytauensis* Bensh, *D. pomposa* Sjom., *D. robusta confinis* Sjom., *D. insignis* Jagof., *D. cybea* Sjom., *Praepseudofusulina immemorata* Malkov., *Pr. prolata* Ketat, *Pr. propria* I. Tcher., *Pseudofusulina paraanderssoni* Raus. f. *acutiterminalis, Ps. pseudoanderssoni* Sjom., *Ps. kireevae* Sjom., *Ps.* ex gr. *paragregaria* Raus., *Rugosofusulina stabilis restricta* Sjom., *Quasifusulina cayeuxi* (Deprat), and *Q. longissima* (Moell.). The appearance of this fusulinid assemblage allowed us to locate a boundary between the Noginsk and Melekhovo horizons with certainty and to include the deposits containing this assemblage in the Melekhovo Horizon. The upper boundary of this horizon in Hole 5 was conventionally located at the level where the last poorly preserved fusulinids, possibly of the genus *Praepseudofusulina,* were observed in a highly dolomitized fusulinid limestone penetrated at a depth of 297 m. A.N. Reimers determined conodonts at a depth of 294 m there, such as *Streptognathodus nodulinearis* Chern. et Resh. and *S. elongatus* Gunn. It follows that the Melekhovo Horizon has a thickness of 10 m in Hole 5.

The fusulinid assemblage that characterizes the Melekhovo Horizon in Hole 5 is identical to the assemblage known from Bed 36 of the Yablonevyi Ovrag section in the Samarskaya Luka area. This section was offered as a hypostratotype for the Gzhelian Stage (Murav'ev *et al.*, 1983) and as a possible stratotype for the Carboniferous–Permian boundary (Murav'ev *et al.*, 1984). In the Yablonevyi Ovrag section, Bed 36 marks the upper boundary of the *Daixina sokensis* Zone. Higher layers yielded the fusulinid assemblage that is dominated by species of the *Daixina robusta* group and large *Daixina* species with a broad free spiral of the *D. bosbytauensis* (Bensh) type, *Praepseudofusulina*, and also by *Rugosofusulina* with thick axial fillings of the *R. stabilis* (Raus.) type, all typical of the assemblage of the Melekhovo Horizon.

Equivalents of the Melekhovo Horizon can be traced in the South Urals in a stratotype of the Carboniferous–Permian boundary deposits of the *Daixina bosbytauensis–D. robusta* and *Schgwagerina vulgaris–Sch. fusiformis* zones, where Ruzhentsev (1950) found *Pseudofusulina gregaria* Lee and *Ps. paragregaria* Raus. in a locality that he referred to as Exposure 10 along the Aidaralash River. Isakova and Nazarov (1986) determined *Daixina* ex gr. *robusta* and *Pseudofusulina kljasmica* (Sjom.) in the same section. A more abundant and diverse assemblage, bearing a close resemblance to the assemblage of the Melekhovo Horizon in its stratotype locality, was reported by Chuvashov *et al.* (1986) from the Aidaralash section. Evidence in support of this similarity is the presence of common species, such as *Daixina vozghalensis* Raus., *D.* aff. *robusta* Raus., *D. bosbytauensis* (Bensh) and various *Praepseudofusulina, Pseudofusulina*, and *Rugosofusulina*, in the deposits of the Aidaralash section and the Moshachikha and Klyazminskii Gorodok sections.

CONCLUSION

The new horizon proposed here for the Ghzelian Stage and named Melekhovo Horizon is traceable in boreholes and in quarries and outcrops of the Gzhelian Stage stratotype locality and elsewhere. The horizon consists of dolomitized limestones and secondary dolomites. The rocks are cavernous, calcitized, and partially silicified; they include chert layers and nodules fusulinids of the *Daixina bosbytauensis–D. robusta* Zone. The total thickness of the horizon is 13 m. In the holostratotype of the horizon (Melekhovo–Fedotovo quarry), its base is marked by the top of the variegated marl bed (Blue Marl) of the Noginsk Horizon. Its lower boundary is defined by a change in the fusulinid assemblage: the assemblage of the *Daixina sokensis* Zone is replaced here by fusulinids of the *Daixina bosbytauensis–D. robusta* Zone. Therefore, the lower boundary of the horizon is distinctly traceable in the succession of the rocks and correlatable on the basis of fusulinid fauna. Earlier, the assemblage of the *Daixina bosbytauensis–D. robusta* Zone was included in the assemblage of the lower interval of the *Schwagerina vulgaris–Sch. fusiformis* s. l. Zone. It was described extensively in the literature and correlated in numerous sections of the East European platform and Urals.

We failed to ascertain a boundary between the Melekhovo Horizon and the base of the Asselian Stage because we did not find *Daixina bosbytauensis* (Bensh) in the Melekhovo stratotype locality of the Gzhelian Stage, nor did we find a sufficient number of *Schwagerina vulgaris* Scherb., or other species of this group, in the basal beds of the Asselian Stage. For this reason, further work needs be done to determine the upper boundary of the Melekhovo Horizon on the basis of fusulinids and other fauna.

It can be stated in conclusion that the results of this study improved the lithological and paleontological criteria for locating the lower boundary of the Melekhovo Horizon in the Gzhelian Stage but were insufficient for the exact delineation of a boundary between the new and Kholodnyi Log horizons of the Asselian Stage in the region of the Moscow syneclise.

The most important species of the new horizon are presented in Plate 1. The newly found species of the horizon are described in the Appendix.

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APPENDIX

Order Fusulinida

Superfamily Fusulinacea Moeller, 1878

Subfamily Schwagerinidae Dunbar et Henbest, 1930

Genus *Praepseudofusulina* Ketat et Solotukhina, 1984

Species *Praepseudofusulina diligens* Isakova, sp. nov.
Plate 1, Figs. 9 and 10

The name of the species is derived
from the Latin *diligens* meaning distinct.

Holotype: Geological Institute, Russ. Acad. Sci., No 4612/9, Gzhelian Stage, Melekhovo Horizon (*Daixina bosbytauensis–D. robusta* Zone); Kovrov area, head of the Moshachikha ravine, Sample 132.

Description. The shell is fusiform with a convex median area and lateral slopes. The internal two or three whorls are short and ovoid. L : D = 1.7–1.8, L = 3.1–3.5 mm, D = 1.8–1.9 mm. The spiral is coiled more closely in the initial three whorls. The fourth whorl is 0.9–1.3 mm in diameter. The proloculus diameter is 140 μm. The number of whorls is 5 or 6. Diameters of the holotype whorls are (mm) 0.14 for proloculus, 0.20 for whorl 1, 0.34 for whorl 2, 0.54 for whorl 3, 0.88 for whorl 4, 1.43 for whorl 5, and 1.87 for whorl 6. The wall is thin, having a maximum thickness of 50 μm in the outer whorl. Plication is ansiform, insignificant, and low in the outer and distinct and sharp in the inner whorls. Sections across the septa look like rounded arcs, mainly on the sides of the shell, and like narrow faveolate bands at the axial extremity not extending to the shell sides. Chomata of proloculus turn further into hooklike pseudochomata. The aperture is narrow and high in the inner whorl and widens abruptly in the two outer.

Comparison. This species resembles *Praepseudofusulina cara* Dobr. by its plication pattern but differs from it by the shell form and the character of coiling.

Stratigraphic range. Gzhelian Stage, Melekhovo Horizon, *Daixina bosbytauensis–D. robusta* Zone.

Material. Ten cross sections.

Genus *Daixina* Rosovskaya, 1949

Species *Daixina insolita* Isakova, sp. nov.
Plate 1, Figs. 5–8

The name of the species is derived
from the Latin *insolitus* meaning unstable.

Holotype: Geological Institute, Russ. Acad. Sci., No. 4612/4, Asselian Stage, Kholodnyi Log Horizon (*Schwagerina vulgaris–Sch. fusiformis* Zone); Kovrov area, Golyshevo village, Sample 15.

Description. The shell has a spherical form with small mammiform axial extremities in the outer whorl. The internal two or two and a half whorls are spherical, those that follow are abruptly elongated up to a fusiform shape, and only the last whorls have stretched out ends. L : D = 1.5–1.8 in the outer whorl and 2.0–2.5 in the middle ones. The shell is large: L = 5.0–6.2 mm, D = 3.06–3.5 mm. The inner two or three whorls are coiled closely to one another; thereafter, the spiral widens rapidly. The diameter of the fourth whorl ranges between 0.75 and 1.22 mm, and the proloculus diameter, between 70 and 100 μm. The number of the whorls is between 5.5 and 7.5. The wall grows thicker, attaining a thickness of 80 μm in the outer whorl. The plication is intense. The thin septa arcs of subtriangular shape in the cross section occupy half of the height of whorls and are stretched upward on the sides of the shell. Narrow strips of fine-faveolate plexus occur in the axial extremities. Chomata are typical only of the inner, closely coiled turns. The aperture is narrow and poorly expressed.

Variability is manifested in the plication pattern: the plicae change from irregular and high to reticulate and low.

Comparison. There are no species among the *Daixina* group resembling this one.

Stratigraphic range. Gzhelian Stage, Melekhovo Horizon, to the Asselian Stage, Kholodnyi Log Horizon (*Daixina bosbytauensis–D. robusta* and *Schwagerina vulgaris–Sch. fusiformis* zones).

Material. Nine cross sections.

Reviewer A.S. Alekseev

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Correlation of Upper Permian Deposits in the Volga–Urals Region and Kuznetsk Basin

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Abstract—The correlation of marine and terrestrial deposits is one of the most difficult problems of stratigraphy. The Volga–Urals region comprises stratotypes of the Upper Permian stages (Ufimian, Kazanian, Tatarian). A thorough lithologic, paleontologic, and paleomagnetic investigation is used to outline regularities and stages in fauna and flora evolution, to estimate the rank of stratigraphic boundaries, and to trace them in the neighboring areas of the Angaraland—Kuznetsk basin. The correlation is based on the analysis of the Late Paleozoic macroflora from the Volga–Urals region and Kuznetsk basin. As a result, the Ufimian Stage of the stratotype locality is correlated with the Usyatskoe, Starokuznetsk, Mitina, and Kazankova-Markina formations, whereas the Kazanian Stage is assumed to correspond to the Uskat and Leninsk formations, and the Tatarian Stage, to the Gramoteino and Tailugan formations of the Kuznetsk basin. Thus, the Lower–Upper Permian boundary in the Kuznetsk basin is placed at the base of the Usyatskoe Formation.

Key words: *correlation, Volga–Urals region, Kuznetsk basin, Upper Permian, stratotype, macroflora.*

The correlation of Upper Permian deposits is now one of the most crucial problems mostly because of a sharp paleogeographic differentiation at the final stage of the Paleozoic evolution of the Earth. The diverse Late Paleozoic floras from continental deposits of Eurasia, isolation of marine basins with endemic fauna, and lithological variability of coeval sequences deposited in different sedimentation settings are the main difficulties we meet when elaborating local stratigraphic scales and conducting interregional correlation.

During prospecting for oil and gas deposits in the Volga–Urals region and coal deposits in the Kuznetsk basin, lithologic and paleontologic investigations were conducted during recent decades, which made it possible to elaborate local stratigraphic scales reflecting the evolutionary stages of both regions, to trace litho- and biostratigraphic units in them, to perform the interregional correlation of the Upper Permian units in the neighboring Angaraland areas, and to tie them to the standard stratigraphic scale. However, the magnetostratigraphic, mineralogical–petrographical (with the use of the nuclear-magnetic resonance method), and electron-microscopic studies of the recent time furnish an opportunity to obtain new data on systematics and taxonomy of fauna and flora and, hence, to refine the stratigraphic information. This paper is aimed at the analysis and correlation of the Upper Permian floras that existed along the margins of the intracontinental Volga–Urals sea and Kuznetsk basin located in the central part of the huge Angaraland.

ANALYZED MATERIALS ON THE UPPER PERMIAN FLORAS OF THE VOLGA–URALS AND KUZNETSK BASINS

Pioneering studies of fossil plant remains from copper shales of the Urals region were performed in the mideighteenth century by F. Wangenheim von Qualen, A. Brongniart, S. Kutorga, G. Fisher de Waldheim, R. Kidston, and D.T. Gwinn-Vaughan. Later, materials collected by A.A. Stuckenberg and P.I. Krotov, Professors of Kazan University, were studied by E. Eichwald and I.F. Schmalhausen, outstanding paleobotanists. In his monograph *Description of Plant Remains from Artinskian and Permian Deposits*, Schmalhausen drew the conclusion, based on the comparison of these floras, that the Artinskian Stage should be attributed to the Permian, not the Carboniferous, System. In the opinion of M.F. Zalesskii, who published results of his studies in the atlas *Permian Flora of the Uralian Areas of Angaraland* (Zalesskii, 1927), flora of the Kama region "is a classic one for the Permian deposits in general, which were studied in this part of the world probably earlier than in other areas" (Zalesskii, 1918, p. 7). The above-listed authors, however, had at their disposal only limited materials from single localities. A more purposeful sampling of plant remains was conducted in connection with geological mapping (Sheet 109) in the 1930s–1950s. As a result, T.A. Tefanova (1963, 1971) sampled a large collection of fossil plants from the lower Kazanian deposits in the vicinity of Bersut, Vandovka, Kamskie Polyany, Sentyak, Kotlovka, and Tar-

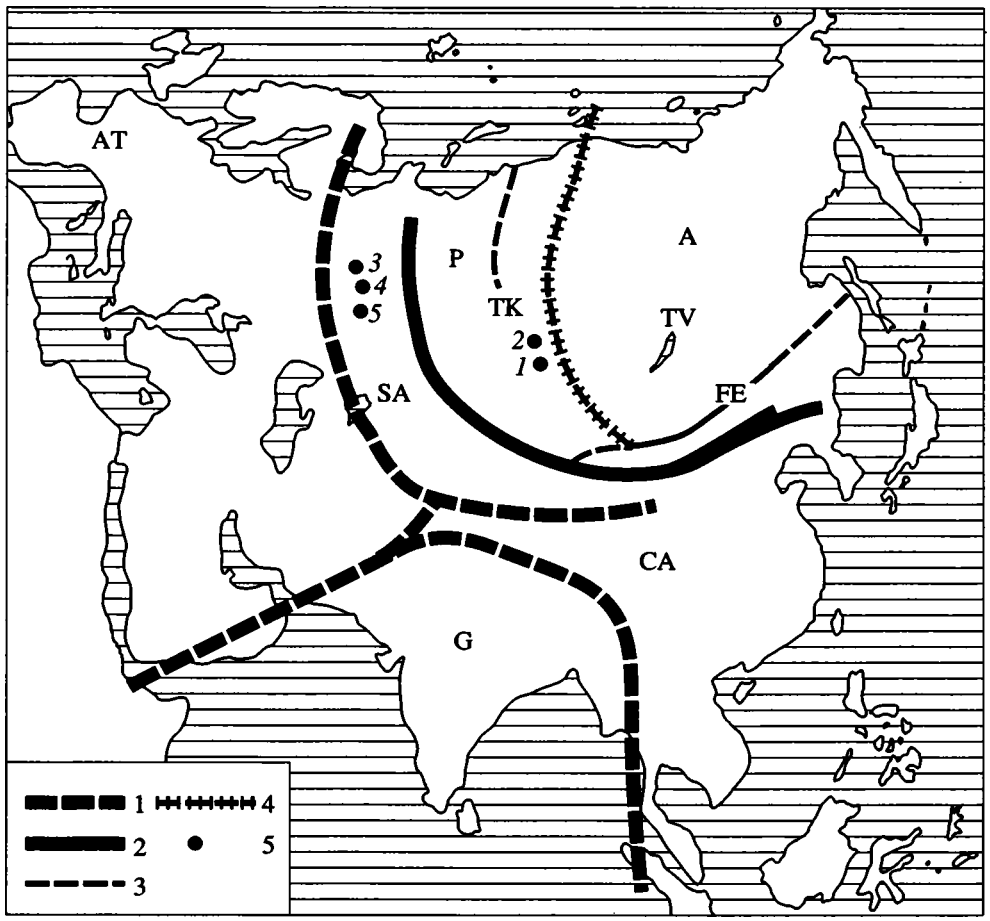

Fig. 1. The phytogeographic zonation of Eurasia in the first half of the Late Permian (after Meyen, 1970): (1) boundary between paleofloral kingdoms; (2) boundary between areas; (3) boundary between provinces; (4) boundary between districts; (5) floral localities; At, Atlantic kingdom; CA, Catasian kingdom; A, Angara kingdom and subordinate zones; SA, Subangara area; FE, Far East area; P, Pechora area; TK, Taimyr–Kuznetsk district; TV, Tungus–Verkhoyansk district; *1*, *2*, Kuznetsk basin; *3*, Kullarovo, Kityak; *4*, Kama River (Elabuga, Tarlovka, Naberezhnye Chelny, Sokolki); *5*, Belebei.

lovka villages situated along the Kama River. The flora list in her papers comprises 32 species of 25 genera. She arrived at the conclusion that two plant assemblages existed in the Kama River lowerstream during the Kazanian time: the older assemblage represented by lepidophytes and the younger one of the coniferous flora with elements of the Mesozoic Ginkgoaceae adapted to a more arid climate. Particularly important for paleogeographic interpretations was the phytogeographic zonation performed by S.V. Meyen. This researcher defined that the East European area "can be equally referred to both the Angaraland and the Euroamerican kingdoms because it includes elements of both regions" (*Paleozoiskie* ..., 1970, p. 154). Later, he named it the Subangara area (Fig. 1). Meyen did not share the opinion of Tefanova on the existence of two humid and arid assemblages in the Kazanian flora. He also argued against characteristics of the Kazanian flora by G.P. Radchenko, who identified it as the "callipteris–odontopterid–coniferous" assemblage. Meyen believed that the appearance of the mesophytic elements in the Kazanian flora is of particular importance and assumed that the East European platform was one of the areas of the mesophytic flora formation.

To correlate sections of the northern part of the Uralian foredeep with the stratotype sections of the Permian stages, scientists from the Komi Division of the Russian Academy of Sciences performed special sampling and study of their fauna and flora. Fefilova (*Korrelyatsiya* ..., 1981) sampled the material from Kazanian deposits along the Sok, Vyatka, and Kama rivers, as well as in the Belebei locality. Following Tefanova, she subdivided the Kazanian Stage into two parts corresponding to lower and upper Kazanian floral assemblages and noted their most characteristic forms, which are common for the Kazanian Stage of the stratotype area and Uralian foredeep. Macrofloral remains from various outcrops of the Kazanian deposits mainly located along the Kama River were studied by Vladimirovich (1984). The Ufimian flora sampled from cores of numerous boreholes in the Sheshma River

basin was studied by A.V. Bogov, who established the narrow stratigraphic range of various species of the *Phylladerma* genus, whereas Vladimiriovich (1982) described flora from the Solikamsk Horizon of the Belaya River valley.

During recent years, I studied almost all outcrops of the Kazanian deposits with floral remains in the Volga and Kama regions, as well as cores from many boreholes (Esaulova, 1986, 1987, 1988). As a result, the detailed paleobotanic characteristics was obtained for the Kazanian Stage from its basal to top layers, including the stratigraphic boundary units of the Ufimian (Sheshma Horizon) and Tatarian (Urzhum Horizon) stages in the marine, transitional, and continental facies. The revision of earlier materials (Esaulova, 1989), along with modern methods allowing preparation of cuticle and phytoleim slides of macrofloral remains and extraction of spores from sporangiophores of permothecas and spore-bearing ferns, substantially supplemented and refined the earlier data.

The Kazanian flora from outcrops along the Vyatka, Sukhona, and Severnaya Dvina rivers was studied by Meyen and Goman'kov (1986), who noted as its distinctive feature the prevalence of peltaspermaceous pteridosperms.

Thus, we have now the sufficiently complete floral characteristics of all three stages of the Upper Permian used to define stages in the macroflora evolution during this period of time; to outline the levels of floral assemblage changes (Esaulova, 1993, 1994); to establish relationships of their boundaries with the lithostratigraphic, biostratigraphic, and paleomagnetic zonations in the Volga–Urals region (Gusev *et al.*, 1993); and to trace them in adjacent regions.

When performing the comparative analysis of the Upper Permian flora of the Volga–Urals region and Upper Paleozoic flora of the Kuznetsk basin, I used various publications (Radchenko, 1934; Neiburg, 1948; Meyen, 1963; Sukhov, 1969; Durante, 1978; Glukhova, 1984; Gorelova, 1970, 1984; *Verkhnii paleozoi* ..., 1988).

CORRELATION OF THE UPPER PERMIAN FLORAS FROM THE VOLGA–URALS REGION AND KUZNETSK BASIN

The problem of correlation of the standard and regional stratigraphic scales is of theoretical and practical importance. It was considered in works by Meyen (1989), Khalfin (1980), Schindewolf (1975), and others. Rejecting the doctrine of catastrophism and accepting the evolutionary character of the fauna and flora development, I believe that this process can be irregular, and the appearance and extinction of fauna and flora can take place over a long period of time, whereas boundaries between floral assemblages might not coincide with boundaries between lithologic units. This, however, does not mean that there are no boundaries of the last type marked by significant changes in the fauna and flora evolution and traceable in space; both fauna and flora actively respond to structural reorganizations of the earth crust. The main purpose of this paper is to analyze regularities in the flora evolution in two neighboring Subangara and Angara areas within the single Angaraland kingdom during the Permian, to outline boundaries in the development of these floras, to determine the rank of the defined boundaries, and to correlate each unit with the standard stratigraphic scale of the East European (Volga–Urals) and local (Kuznetsk) basins.

Paleobotanic characteristics of the Upper Permian deposits in the Volga–Urals region allows the Late Permian flora to be considered as inherited to a certain extent from the "Bardian" (Artinskian–Kungurian) flora. In general, it is characterized by the prevalence of pteridosperms, ferns, and articulates, as well as by the insignificant role of cordaites (Table 1). In the Solikamsk Horizon, leaf remains, lycopod bark, and bryophytes are dominant. The Sheshma Horizon is the first occurrence level of leaves of peltaspermaceous pteridosperms of the *Phylladerma* genus characteristic of the Permian deposits of the Volga–Urals region. The Sok and Volga horizons are marked by the most diverse flora. The Sok Horizon exhibits the mass appearance of various ferns, pteridosperms, abundant articulates, and conifers; large leaves of *Odontopteris, Zamiopteris, Psygmophyllum, Codaites,* and, less commonly, small *Rufloria* are also characteristic. The flora of the Volga Horizon retains to a great extent the older appearance, though it is distinct because of the first occurrence of mesophytic elements such as *Taeniopteris*. The flora of the Urzhum Horizon is substantially impoverished and includes elements of vegetation from the Sok and Volga horizons.

Significant vegetation renewal occurred in the Severnaya Dvina period of time, which was marked by the first occurrence and prevalence of representatives of the *Tatarina* flora of mesophytic appearance. *Rhaphidopteris* and rare *Cordaites* are present here; however, ruflorians disappear, and bryophytes and ferns play a subordinate role. The flora of the Vyatka Horizon is similar in general to that from the Severnaya Dvina Horizon but lacks *Cordaites*.

On this basis, I assumed that three floral assemblages existed during the Permian: (1) *Psygmophyllum* or the "Bardian" Artinskian–Kungurian flora; (2) *Phylloderma* flora (Kama or Sheshma–Kazan–Urzhum assemblage, whose key taxa are represented by the Kazanian flora); and (3) *Tatarina* flora of the late Tatarian stage (Esaulova, 1993). Most distinct levels marked by vegetation changes coincide with the boundary between the Solikamsk and Sheshma horizons within the Ufimian Stage and with the boundary between the Urzhum and Severnaya Dvina horizons within the Tatarian Stage.

The change in depositional environments in the Volga–Urals region during the Permian was also

Table 1. Upper Permian floral assemblages of the Volga–Urals region (typical taxa are underlined)

System	Series	Stage	Substage	Horizon	Paleobotanic characteristics
Permian	Upper	Tatarian	upper	Vyatka	*Pelliotilites tataricus, Protosphagnum nervatum, Rhizinigerites neuburgae, Muscites* sp., *Fasciastomia delicata,* articulate stems, *Fefilopteris papilosa, Pecopteris* sp., *Peltaspermopsis buevichiae, Tatarina conspicua, T. pinnata, Tatarina olferievii, T. mira, Saplingocarpus bicornutus, S. variabilis, Phylladoderma (Aequistomia) aequalis, P.(A.) annulata, P. (A.) tatarica, Doliosomia krassilovii, Rhaphidopteris kiuntzelae, R. antiqua, Permotheca striatifera, P. vesicasporoides, P. vittatinifera, Pursongia beloussovae, Sphenarion ?* sp., *Sashinia aristovensis, S. borealis, Dvinostrobus sagittalis, Quadrocladus dvinensis, Q. borealis, Geinitzia subangarica, Pseudovoltzia ?* sp., *Arisada densa.* (Goman'kov and Meyen, 1986)
				Severnaya Dvina	*Protosphagnum nervatum, Fasciastomia delicata,* articulate stems, *Fefilopteris pilosa, Pecopteris* sp., *Peltaspermopsis buevichae, Tatarina olferievii, T. conspiqua, Saplingocarpus bicornutus, S. variabilis, Phylladoderma (Aaquestomia) annulata, P. (A.) rastorguevii, P. (A.) trichophora, Rhaphidopteris* sp., *Permotheca striatifera, P. vesicasporoides, Pursongia cf. amalitzkii, P. beloussovae, Cordaties clercii, Dvinostrobus sagittalis, Quadrocladus dschweitzerii, Geinitzina* sp., *Arisada densa.* (Goman'kov and Meyen, 1986)
			lower	Urzhum	*Paracalamites friqidus, Annularia* sp., *Signacularia noinskii, Sphenophyllum (Tichvinskia) stoukenbergii, Pecopteris* sp., *Rhaphidopteris praecursoria, Phylladoderma* sp., *Permotheca sardykensis, P. vesicasporoides, Taeniopteris* sp., *Tatarina* sp., *Nucicarpus minutus.* (Esaulova, 1986, 1987)
		Kazanian	upper	Volga	*Singacularia noinskii, Viatscheslaviophyllum kamiense, Paracalamites frigidus, P. kutorgae, Phyllotheca* sp., *Annularia pseudostellata, Tschernovia striata, Annulina neuburgiana, Sphenophyllum (Tichvinshia) stoukenbergii, Prynadoepteris* sp., *Pecopteris anthriscifolia, Odontopteris rossica, O. tatarica, Rhaphidopteris* sp., *Comia macrophylla, Phylladoderma volgensis, Taeniopteris eckardti, Pseudovoltzia? cornuta, Permotheca sardykensis, Peltaspermum* sp., "a," *Nicicarpus minutus, Sylvella heteromorpha.* (Esaulova, 1986, 1987)
			lower	Sok	*Signacularia noinskii, Viatscheslaviophyllum kamiense, Paracalamites frigidus, P. decoratus, P. similis, P. kutorgae, Paracalamostachys* sp., *Phyllotheca deliquescens, Annularia stellatoides, Sphenophyllum (Tichvinskia) stoukenbergii, Asterotheca* sp., *Oligocarpia* sp., *Todites* sp., *Danaeites (=Orthotheca)* sp., *D.(=O.) petschorica, Pecopteris anthriscifolia, P. helenaeana, Lobatopteris* sp., *Brongniarthites salicifolius, Compsopteris* sp., *Callipteris* sp., *Odontopteris rossica, O. tatarica, Rhaphidopteris* sp., *Psygmophyllum nesterenkoi, Ps. expansum, Zamiopteris* sp., *Pursongia* sp., *Phylladoderma meridionalis, P. sentjakensis, Rhipidopsis* sp., *R. ginkgoides, Taeniopteris tingii, Rufloria synensis, Cordaites candalepensis, Pseudovoltzia? cornuta, Quadrocladus sp. Hydropterangium* sp., *Permotheca vesicasporoides, P. sardykensis, Peltaspermum* sp., *Samaropsis typ. vorcutana, Nucicarpus piniformis, N. sentjakensis, Carpolithes* sp., *Entsovia rarisulkata, Timanostrobus?* sp., *Glottophyllum* sp., *Cordaicarpus petrikensis* (Esaulova, 1986, 1987; *Korrelyatsiya* ..., 1981)
		Ufimian	upper	Sheshma	*Singacularia noinskii, Viatscheslavia vorcutensis (?), Viatscheslaviophyllum kamiense, Paracalamites frigidus, P. cf similis, Pecopteris* sp., *Psygmophyllum expansum, Carpolithes* sp., *Samaropsis* sp., *Cordaites* sp., *Phylladoderma spinosa, Brongniarthites salicifolius, Prynadaeopteris anthriscifolius, Callipteris tenuinervia, Odontopteris* sp., *Nucicarpus piniformis.* (Esaulova, 1986, 1987; *Korrelyatsiya* ..., 1981)
			lower	Solikamsk	*Jntia variabilis, J. vermicularis, Viatscheslavia vorcutensis, Viatscheslaviophyllum kamience, Paracalamites frigidus, Pecopteris* sp., *Odontopteris artipinnata, Comia rarinervia, Syniopteris expansa, Rufloria ensiformis, Entsovia rarisulcata, Permotheca sardykensis, Samaropsis vorcutana, S. extensa, S. triquetra, S. elegans, Sylvella* sp., conifer fragments. (Vladimirovich, 1982; *Korrelyatsiya* ..., 1981; Esaulova, 1986, 1987)

Table 2. Floral assemblages from the Kol'chugino Group of the Kuznetsk basin (typical taxa are underlined)

Regional stratigraphic units of the Kuznetsk basin (*Verkhnii paleozoi* ..., 1988)			
Group	Subgroup	Formation	Paleobotanic characteristics
Kol'chugino	Erunakovo	Tailugan	*Annularis (?) grandifolia, A.(?) jerunakovensis, Prynadaeopteris venusta, Pecopteris tajmyrensis, Todites evenkiensis, Comia maganakovkiensis, Zamiopteris tajluganensis, Cordaites minutus, Lepeophyllum tajluganensis, Ginkgoites plotnikovaensis, Phipi-dopsis palmata, Tomia oblonga, Samaropsis erunakovensis, Yavorskia mungatica.*
		Gramoteino	*Jniopteris sibirica, Rhipidopsis palmata, Ginkgoides plotnikovoensis, Tychtopteris cuneata,* abundant *Cordaites clercii.*
		Leninsk	*Koretrophyllites tomiensis, Phyllotheca turnaensis, Annularia lanceolata, Annulina kuschejakovensis, Sphenophyllum tomiense, Prynadaeopteris leninskiensis, Glottophyllum cuneatum, Tychtopteris cuneata, Rufloria mitinaensis, R. olzerassica, Cordaites insignis, C. clercii, C. mediocris, Samaropsis irregularis, S. erunakovensis, S. kureikaensis, Sylvella (?) dubia, Comia osinovskiensis.*
	Il'inskoe	Uskat	*Phynadaeopteris nataliana, Callipteris lobatus, C. altaica, C. zeilleri, C. adzvensis, Samaropsis grjasevii, Cordaicarpus petrikensis, Rufloria minuta, Listrophyllum uscatense, Koretrophyllites grandis, Nephropsis schmalhausenii, Paracalamites iljnskiensis, Tungussocarpus tychtensis, T. elongatus, Samaropsis erunakovensis, S. irregularis, Sylvella dubia, Schizoneura sibirica, Listrophyllum uscatense, Polyssaevia spinalifolia, Cordaites kuznetskiana, C. mediocris, Rufloria olzerassica, R. mitinaensis, R. brevifolia.*
		Kazankova–Markina	*Cordaties gracilensis, C. gorelovae, Rufloria brevifolia, R. motimaensis, Glottophyllum elongatum, G. karpovii, Tychtopteris primitiva, Polyssaevia, Tungussocarpus tychtensis, T. elongatus, Samaropsis erunakovensis, S. irregularis, Condamajella gorelovae.*
	Kuznetsk	Mitina	*Phyllopitys (?) sessilifolia, Phyllotheca turnaensis, Annulina iljinskiensis, Paracalamites communis, Zamiopteris crassinervis, Nephropsis grandis, N. lampadiformis, Callipteris ivancevia, Cordaicarpus petrikensis, Tungussocarpus elongatus.*
		Starokuznetsk	*Annularia planifolia, Cordaites latifolius, C. gracillentus, C. minax, Rufloria derzavinii, R. olzerassica, Crassinervia peltifromis, Cr. ivancevia, Cr. pentagonata, Pecopteris pseudomartia, Callipteris ivancevia, Samaropsis pseudotriquetra, S. trapeziformis, Sylvella dubia, Tungussocarpus elongatus.*

reflected in the development of freshwater bivalves (Fig. 2). The most pronounced changes in this group coincide with the boundary between the Urzhum and Severnaya Dvina horizons (Gusev, 1993). It is not inconceivable that these faunal and floral changes were coeval with the inversion of the magnetic field at the boundary between the Kiama and Illawara hyperzones or with the reorganization of structural patterns of the earth crust; they were of a global scale and could be established in the neighboring region.

A paleobotanic substantiation of stratigraphic units in the Kuznetsk basin was performed by Gorlova (1970, 1984) and Durante (1978). These authors provided the thorough floral characteristics of all formations of the Kol'chugino Group (*Verkhnii paleozoi* ..., 1988) and defined several guide forms (Table 2) and correlative floral levels based on cordaites and other macrofloral taxa (Fig. 3). On the whole, the flora from the Kol'chugino Group of the Kuznetsk basin is characterized by abundant and diverse cordaites and thus significantly differs from the Upper Permian flora of the Volga–Urals region. The thoroughly studied anatomical structure of cordaites shows their principal significance for subdivision and correlation of the section within the entire Angaraland kingdom (*Osnovnye cherty* ..., 1984, Table 5). The flora of the Starokuznetsk Horizon is of a mixed type. Its *Callipteris* and "small-leaved taxa appear and coexist with the last large-leaved *Cordaites* and *Rufloria* of the Balakhonka flora," thus indicating, in the opinion of Gorelova (*Verkhnii paleozoi* ..., 1988, p. 24), the commencement of a new, rufloria–cordaites–callipterid stage of floral evolution. The Mitina flora (lacking the Balakhonka elements) was already typical of the Kol'chugino time.

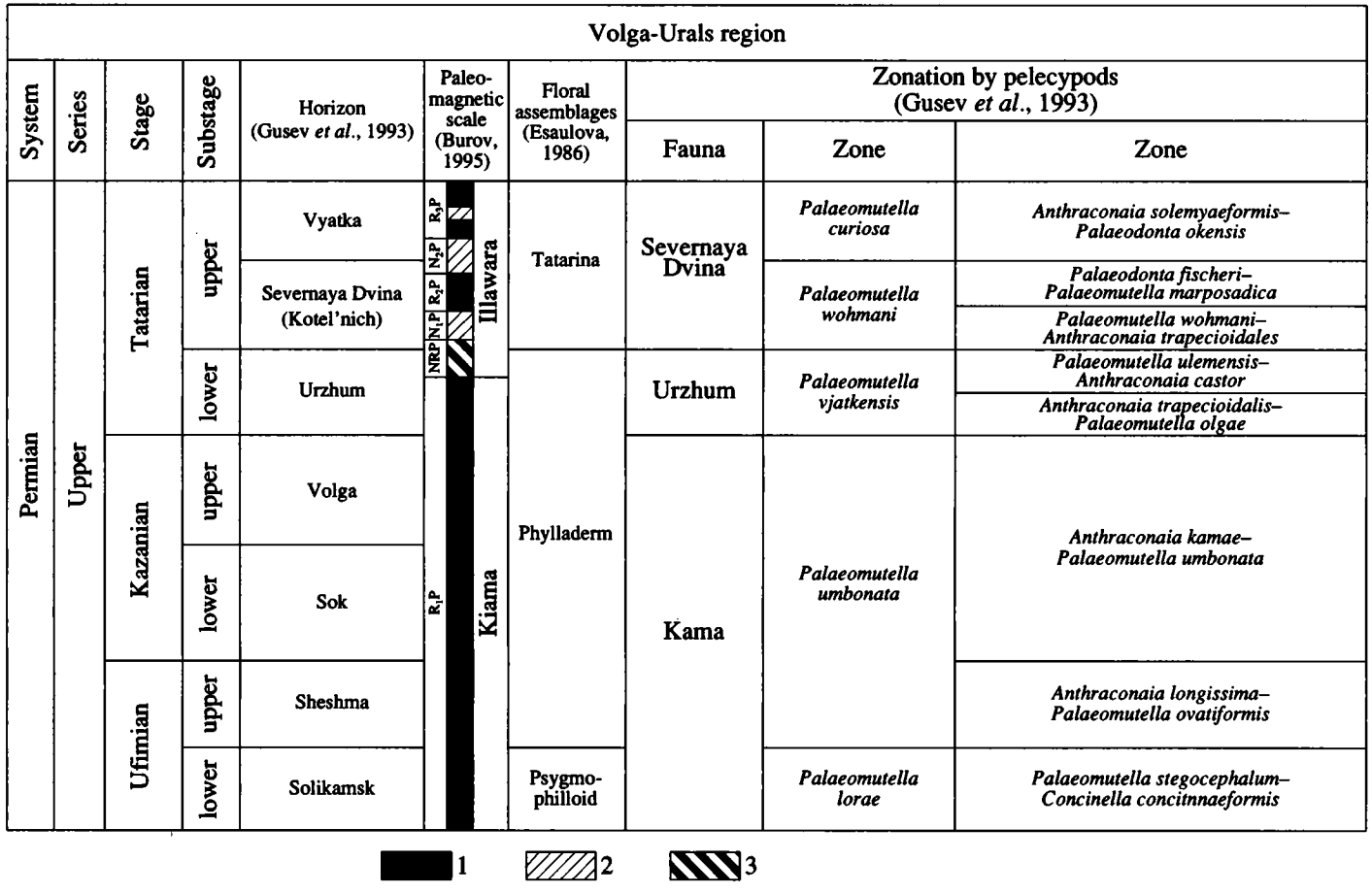

Fig. 2. The Upper Permian stratigraphic scheme of the Volga–Urals region: (1) zone of normal magnetic polarity; (2) zone of reversed magnetic polarity; (3) zone with variable magnetic polarity.

Kuznetsk basin						Standard stratigraphic scale		
Group	Subgroup	Formation	Correlation floral assemblages (Durante, 1978)	Floral correlation beds (*Verkhnii paleozoi* ..., 1988)	Local zones by bivalve mollusks (*Verkhnii paleozoi* ..., 1988)	System	Series	Stage
Kol'chugino P_2	Erunakovo	Tailugan	*Sulcate cordaites*	*Yavorskya mungatica*	*Concinella clara* *Tajluganica tajluganensis*	Permian	Upper	Tatarian
		Gramoteino		*Cordaites clercii, Rhipidopsis palmata*	? *Adontella elegantula*			
		Leninsk	*Ruflorians–sulcate*	*Pecopteris leninskiensis, Phyllotheca turnaensis*	*Anadontella subraphillipsis–Tersiella certa*			
	Il'inskoe	Uskat	IV *Rufloria-Cordaites*	*Rufloria minuta*	*Anadontella molchanovi–Brusiella simplex*			Kazanian
		Kazankova-Markina		*Crassinervia elliptica, Rufloria brevifolia*	*Anadontella iljinskiensis, Palaeodonta pseudolongissima*			
	Kuznetsk	Mitina	III *Rufloria–Cordaites*	*Callipteris ivancevia*	*Palaeomutella markini*			Ufimian
		Starokuznetsk	II *Rufloria–Cordaites*	*Comia osipovskiensis*	? *Augea elliptica*			
Balakhonka P_1	Upper Balakhonka	Usyatskoe		*Salairia longifolia*	*Abiella ussovi–Procopievskia sibirica*		Lower	Kungurian
		Kemerovo		*Sylvella, Skokia*	*Procopievskia gigantea–P. sygmoidae*			Artinskian
		Ishanova	I *Rufloria–Cordaites*	*Zamiopteris glossopteroides, Pursongia sibirica*	Not defined			
		Promezhutochnaya	III *Rufloria*	*Evenkiella schortonensis*	*Kemeroviella tenuistriata* *Mrasiellina umbonata*			Sakmarian-Asselian

Fig. 3. The Upper Paleozoic scheme of the Kusnetsk basin (*Verkhnii paleozoi*..., 1988).

The Mitina flora of the Kazankova–Markina Horizon is dominated by small-leaved *Cordaites, Rufloria,* Bryophyta, and seeds. The flora of the Uskat Horizon is diverse and abundant; in addition to abundant *Cordaites*, ferns and callipterids appear here. The Leninsk Horizon is characterized by a similar floral diversity, contains the larger-leaved forms, and marks the appearance of mesophytic elements. Flora of the Gramoteino Horizon contains diverse callipterids, along with abundant *Rhidiopsis, Ginkgoides,* and *Iniopteris*, but is almost lacking *Rufloria.* The floral assemblage of the Tailugan Horizon is dominated by various ferns and pteridosperms; cycadophytes, ginkgoales, and "dwarfish" sulcial cordaites also occur here. Gorelova and Betekhtina (*Verkhnii paleozoi* ..., 1988) noted an almost complete coincidence of levels of faunal and floral changes in the Balakhonka time; however, such levels of the Kol'chugino time are diachronous and complicate the problem of section subdivision and correlation in the basin.

The question about boundaries between subseries and series of the Permian System in the Kuznetsk basin was repeatedly raised. Stressing the mixed character of the fauna from the Starokuznetsk Horizon, Khalfin

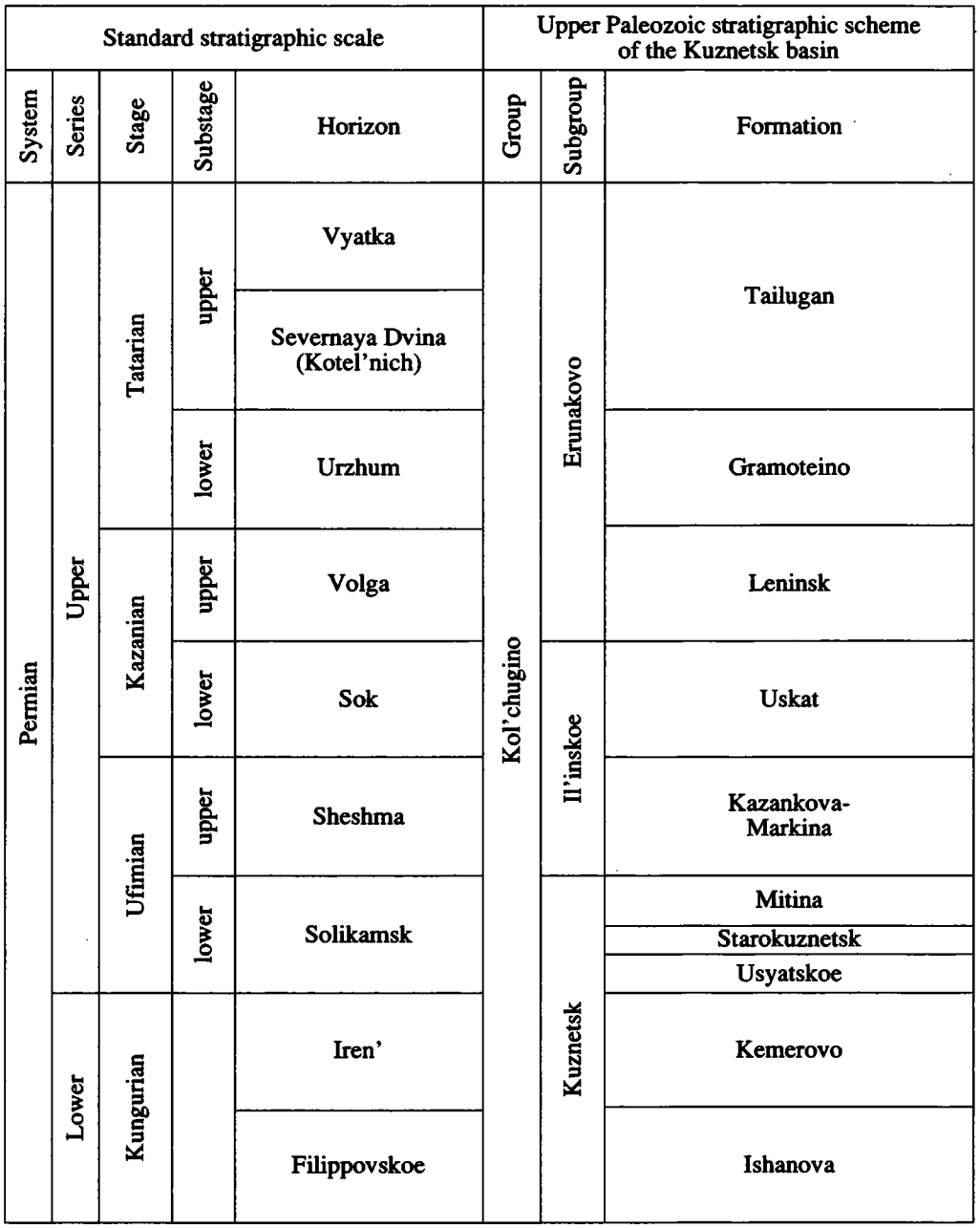

Fig. 4. Correlation of the Upper Paleozoic units of the Kuznetsk basin with the standard stratigraphic scheme proposed in this paper.

considered this feature to be recurrent and termed the fossils as "the second Alakaevo fauna," whereas Tokareva referred it to the particular stage of the Kol'chugino fauna formation. Betekhtina noted the substantial variability of bivalves and their high diversity in the Usyatskoe Horizon, explaining this by a sharp increase in the basin salinity due to "probable ingression of water masses of a different salinity" (*Verkhnii paleozoi* ..., 1988, p. 23). Gusev (personal communication) also does not rule out the connection of two basins in the Late Permian, treating the species *Paleomutella markini* (a taxon that is present in the synonymous zone in the Mitina Horizon) as common to both of them. In the Volga–Urals region, this form is widespread in the Solikamsk Horizon. In my opinion, it is incorrect to draw the Lower–Upper Permian boundary in the Kuznetsk basin based on the appearance of *Callipteris,* because callipterids are also present in the Bardian (Artinskian–Kungurian) flora of the Urals. The mass appearance and wide distribution of ferns, pteridosperms, and mesophytic elements in the Uskat and Leninsk time represent a principal stage in the evolution of the Kuznetsk basin flora. In the Volga–Urals region, similar changes in floral assemblages occurred

in the Sok and Volga times of the Kazanian Age. Therefore, it is more logical, in my opinion, to correlate the Uskat and Leninsk formations with the Kazanian Stage of the stratotype region. The second important level in the Volga–Urals region is marked by the extinction (dying off) of cordaites, as the Urzhum Horizon lacks ruflorians, and, in the Severnaya Dvina Horizon, cordaites are missing. In the Kuznetsk basin, ruflorians disappear at the level of the Gramoteino Formation, and in the Tailugan Formation large-leaved cordaites are replaced by dwarfish forms. The important boundary in the bivalve evolution coincides with the lower boundary of the Taigulan assemblage, distinct because of a high diversity, appearance of new species and genera, and small dimensions of specimens.

Comparing the microstructure of cordaites from the Kuznetsk basin and Urals region, L.V. Glukhova noted that "in general, the assemblage of Kungurian ruflorians is similar to the Ishanova assemblage" (Glukhova, 1984, p. 121). Based on the results of study of the "Bardian" flora from the Urals, Naugol'nykh (1993, p. 61) argues that "according to several common taxa, the Kungurian Stage is well correlative with the Ishanova Horizon of the Kuznetsk basin."

Based on the evolutionary stages of flora from the Angara and Subangara areas, it is valid, in my view, to divide the Permian flora of the Kuznetsk basin into three major floral assemblages: (1) the *Cordaites–Zamiopteris* assemblage that comprises the intermediate, Ishanova, and Kemerovo floras; (2) the *Papillophyllites–Feonia* assemblage of the Uskat and Leninsk floras; and (3) the *Javorskyia* assemblage corresponding to the Tailugan flora. The first assemblage, in my opinion, is an analogue of the psygmophilloid ("Bardian") flora, whereas the second and third assemblages represent the phylladerm (Kazanian) and the *Tatarina* (late Tatarian) floras, respectively.

Thus, the version of correlation of the standard stratigraphic scale of the Volga–Urals region with the local scale of the Kuznetsk basin proposed in Fig. 4 seems more logical.

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Stratigraphy and the Neogene–Quaternary Sedimentation Environment in the Northwestern Shelf and Coast of the Tartar Strait

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Abstract—The systematic analysis of geological, geomorphological, lithological, and paleontological data is used along with results of geochronological studies (C^{14}, TL, and AA dating) to detail the Upper Cenozoic stratigraphy of shelf deposits and to detect their sedimentation environment. As inferred, the sedimentary cover of the shelf comprises a thick sequence of Miocene–Pliocene rocks. The oldest sediments of the terminal Miocene (presumably older than 5 Ma) were deposited here under conditions of a rather cold climate. The lower Pliocene deposits are rich in pollen of the Turgai and subtropical floras, and their beds also yielded molluscan shells, whose stratigraphic range corresponds well to the age interval of 5.1–4.3 Ma estimated on the basis of the diatom flora. The upper Pliocene sediments bear boreal and arctic mollusks associated with periglacial plant remains. *Neodenticula koizumi* from these sediments suggests that their age interval is 4.3–3.0 Ma. The Eopleistocene strata include two members, one corresponding to the initial cool epoch of this stratigraphic interval and another correlative with the warm episode of the Jaramillo Subchron (1.3–0.79 Ma). The respective diatom assemblage and silicoflagellate *Mesocena elliptica* substantiate the correctness of the last age range. The Quaternary System of marine sediments in the region is divided into the lower–middle Pleistocene beds and two higher units, one corresponding to the lower–middle part of the upper Pleistocene and another to the Holocene age.

Key words: *stratigraphy, paleogeography, shelf, Pliocene, Pleistocene, Holocene, transgression, regression.*

Despite a certain progress in exploring the Quaternary shelf deposits, they still inadequately studied, especially the Pliocene and Lower Quaternary marine sediments in the Far East are. This concerns most of all the Neogene unconsolidated deposits. At present, investigations in this field are at the stage of obtaining and collecting data, and the available conclusions remain doubtful. Materials from drilling in the western shelf of the Tartar Strait (operations were performed by the drilling vessel "Geolog Primor'ya" of the Joint-Stock Company "Dal'morgeologiya" in 1991) considerably supplement information on the shelf evolution during the Pliocene and afford a possibility to elucidate the paleogeographic features of the area at that time. These and respective paleontological data were used to refine and modify the available stratigraphic schemes of the region (Kuz'mina *et al.*, 1994; Kuz'mina and Taldenkova, 1994).

The main conclusions of this work are based on data obtained from drilling Sites 1, 2, 5, 6, and 7 and from the prospect trench 110 at the coastal terrace (Fig. 1). Description of cores and selection of samples were conducted during the drilling process.

The section of deposits in the region consists of diverse facies of marine, coastal-marine, and lagoonal sediments. We failed to detect any traces of terrestrial sediments here. Analyzing the geological and geomorphological features of the region along with the lithology of the sediments, and using data on terrigenous, authigenic, and clay minerals along with distribution patterns of spores, pollen, diatoms, foraminifers, and mollusks, we recognized 13 stratigraphic beds dated by geochronological methods, such as ^{14}C, thermoluminescence (TL), and amino acid (AA) dating. All these units correspond to certain stages of sedimentation in the western Tartar Strait during the Neogene–Quaternary time (Figs. 2, 3).

Bed I (Figs. 2, 3). The bed consists of alternating silt and sand layers, the latter bearing rare well-rounded rock fragments along with abundant whole or damaged molluscan shells. The composition of the heavy mineral fraction from the sediments points to a complex history of their accumulation. The sediments display erosion horizons reflecting breaks in sedimentation, lacking authigenic components, and mainly consisting of rock-forming and accessory minerals. The authi-

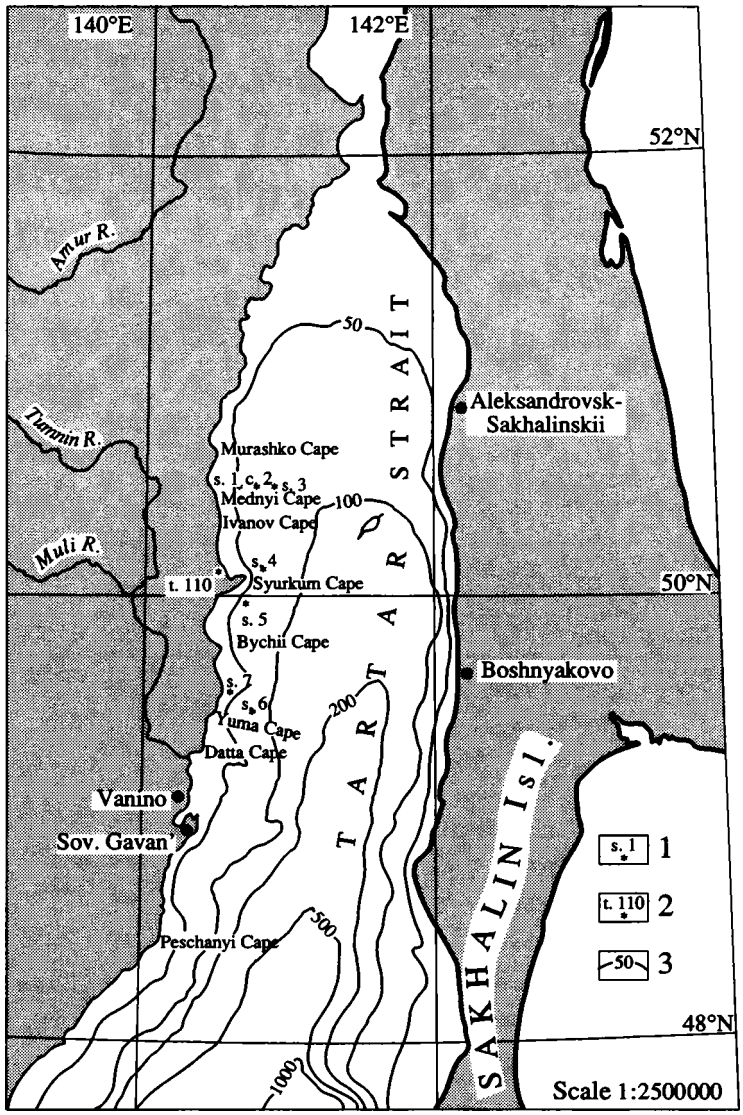

Fig. 1. Geographic scheme of the studied region: (1) Drilling site in the shelf; (2) prospect trench; (3) isobath, meters.

genic components of the bed are chiefly represented by Fe-sulfides associated in its basal and upper portions with Fe- and Mn-oxides and hydroxides (mainly with limonite). Some clay minerals, for instance, the fragmented and often rounded kaolinite crystals, suggest a somewhat unstable regime of sedimentation.

Fossil mollusks of diverse taxonomic composition are mostly buried in situ. Changes in the abundance and composition of their shells in separate layers point to certain fluctuations of the basin depth. Shells of *Liocyma fluctuosa* are preponderant in the lower part of the bed. The most interesting molluscan assemblage from the upper part includes shells of extinct taxa, such as *Fortipecten* sp. and *Chlamys cosibensis*. These taxa were characteristic of waters near Sakhalin Island in the period of the latest Miocene to the terminal early Pliocene, and then they ceased to live (*Atlas* ..., 1968; *Biofatsial'nye* ..., 1974; Kafanov, 1986). The shallow-water thermophilic genera *Fortipecten* and *Anadara* present among them and associated with abundant calcareous nereites suggest a rather warm environment in the past shelf zone and presumably are indicators of the incipient early Pliocene transgression.

The bed yielded many diverse foraminifers: *Porosorotalia clarki* (Voloshinova), *Ammonia nipponica* (Asano), *A. japonica* (Hada), *Saidovella katangliensis* (Voloshinova et Borovleva), *Pseudoelphidiella problematica* (Voloshinova), and *Elphidiella oregonensis* (Cushman et Grant). Forms with a complicated ornamentation representing the family Elphidiidae and species of the genus *Ammonia* again suggest warm and shallow conditions in the past basin. The *Ammonia* species are able to multiply when the water temperature is at about 18°C during at least a short season of the year;

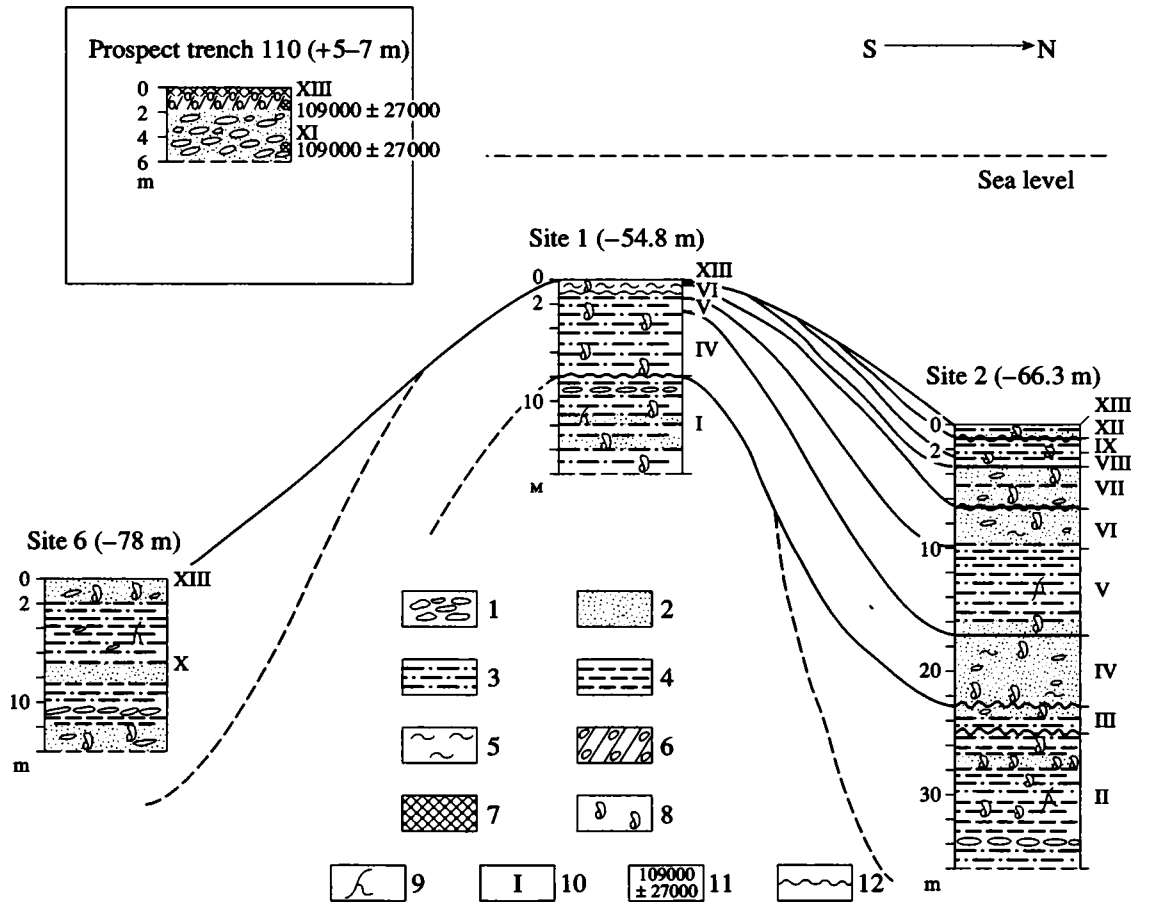

Fig. 2. Correlation scheme of shelf and coastal deposits of the western Tartar Strait: (1) pebbles; (2) sand; (3) silt; (4) pelite; (5) ooze; (6) loam with boulders; (7) soil; (8) molluscan shells; (9) plant remains; (10) bed numbers (I, Ne_I^3 ; from II to IV, Ne_{II}^1 ; V and VI, Ne_{II}^2 ; VII and VIII, EQ; IX, Q_{I-II}; X, $Q_{II}^{2,4}$; XI, Q_{III}^1 ; XII, Q_{III}; XIII, Q_{IV}); (11) TL age data; (12) break in sedimentation.

otherwise they cease their reproduction, as is inferred for the late Miocene stage of foraminifer evolution in Sakhalin (Voloshinova *et al.*, 1970).

Spore–pollen data for drilling Site 1 reveal an alternation of cooling and warming phases in the plant evolution. Three phases of cooling are marked by palynoflora with predominant pollen of *Larix* (34–68%) and less abundant pollen of ancient *Picea* species (11–30%). Flora of this type also include pollen of cold-resistant fruticose and fruticulose plants, such as *Betula* sect. Fruticosae (4–15%), *Betula* sect. Nanae (3–20%), and *Alnaster* (up to 4%), whereas specimens of thermophilic plants are missing in its composition. Intervals corresponding to warmer climatic phases are richest in *Picea* pollen (17–71%) associated with that of old *Pinus* taxa (10–34%), *Abies* (up to 8%), *Tsuga diversifolia* (Max), and others (up to 3%), though pollen of broad-leaved plants is extremely rare here.

Thus, vegetation in the coastal zone changed three times from larch forests with subordinate spruce trees and local areas of fruticulose tundra, which were typical of cooler phases, to spruce taiga, which descended from the Sikhote Alin Ridge down to the sea shore during the relatively warm periods. This alternation of cooler and warmer climatic phases developed simultaneously with the general gradual cooling of the climate at the end of the late Miocene (Akhmet'ev, 1973; Bolotnikova, 1979; Sato, 1963).

We measured the amino acid ratios in shells of the genus *Mya* from the bed in order to evaluate the age of the sediments. The maximum values of the D-alloisoleucine/L-isoleucine ratio turned out to be close to 1.33 (in the free fraction). This value approaching the equilibrium (Miller and Brigham-Grette, 1989) supports our opinion that the sediments of Bed 1 are the oldest among those studied in this work.

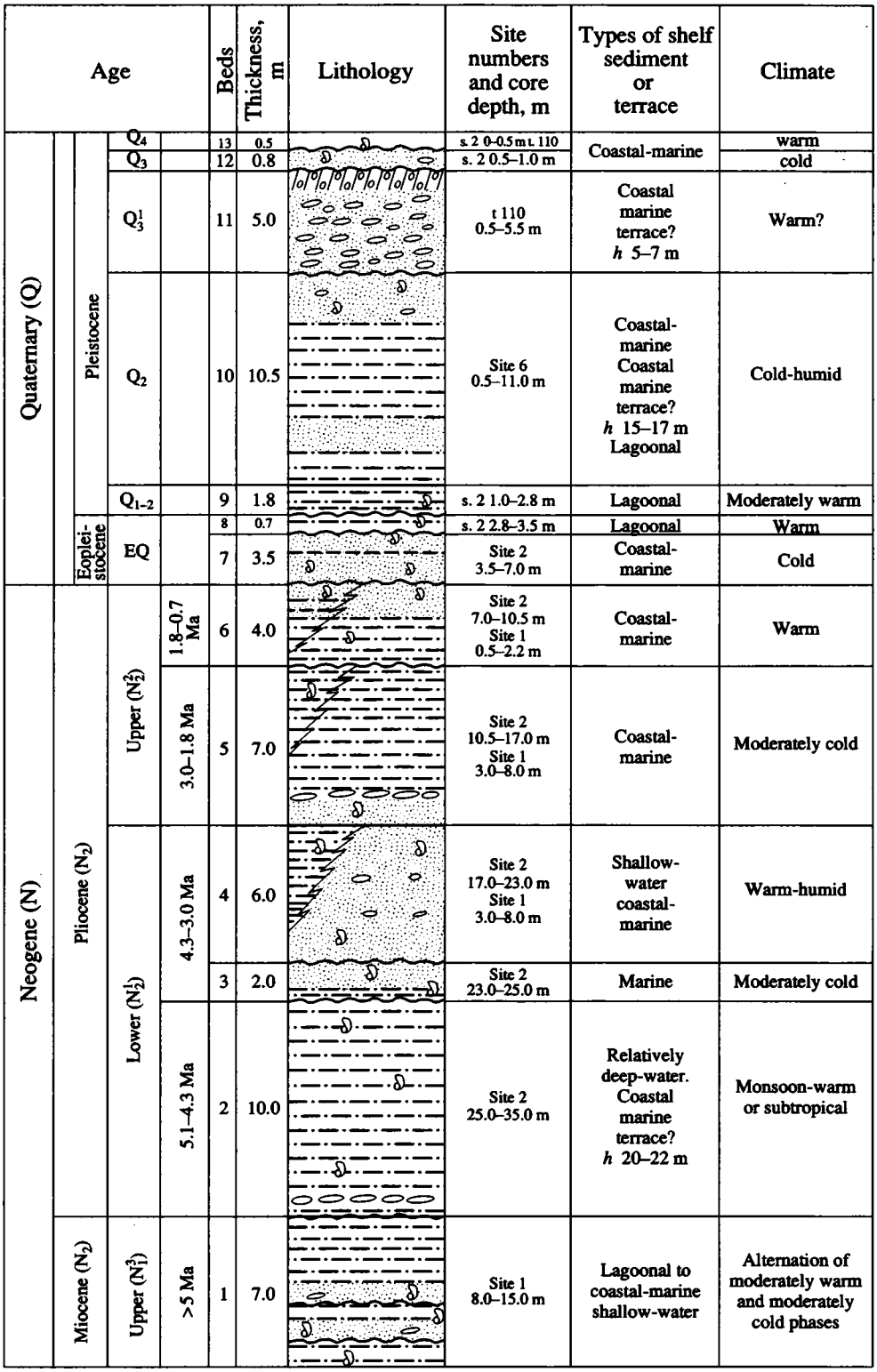

Fig. 3. Composite section of Neogene–Quaternary deposits in the western Tartar Strait (symbols for lithology as same in Fig. 2).

Bed II (Figs. 2, 3). The bed is composed of layered compact silt and pelite layers enclosing pebbles and shell detritus; its thickness is about 10 m. The heavy fraction of rocks mostly consists of authigenic minerals (up to 93%), and pyrite representing one of them is concentrated in the basal interval. Higher in the bed, pyrite (up to 40%) coexists with Fe–Mn-compounds (up to 50%) and occasional hematite. The reduction state in

basal rocks grades upsection into the neutral and oxidation conditions, and the rocks gradually turn into marine deposits of a relatively deep shelf.

As the region of study is located close to the boundary between the Arctic-Boreal and Tropical biogeographic provinces, the diatom stratigraphy important in this case is based on the respective datum levels substantiated at present by a considerable amount of information (Koizumi, 1985). The diatom distribution in sediments is also useful for interpretation of the past sedimentation environments (Jouse, 1961; Pushkar', 1979).

The diatom assemblage from Bed II is referred to the early Pliocene (5.1–4.4 Ma). Species of stratigraphic significance at this level are *Thalassiosira oestruppi* (Ost.) Proch.-Lavr., *T. jouseae* Akiba, *Neodenticula kamtschatica* (Zabelina, Akiba et Yanasigawa), and *Cosmiodiscus insignis* Jouse. Other specimens of the assemblage also belong to marine neritic species of the arctic-boreal type widespread in the lower half of the Pliocene. In addition, the assemblage includes rare valves of the thick-walled Miocene species *Pinnularia viridis* (Nitz.), which is a representative of the freshwater benthos that is very dissolution-resistant and presumably redeposited by rivers. The general aspect of the assemblage is characteristic of the open and relatively deep settings in the shelf zone.

Molluscan shells from the lowermost part of the bed are extremely weathered and easily breakable. They seem to be rather old and suggest the diagenetic alteration of host sediments. At higher levels, the degree of weathering decreases, and the occurrence mode of the shells excludes their redeposition. The molluscan assemblage consists here of *Liocyma fluctuosa, Mactra* sp., *Spisula* sp., *Clinocardium* sp., *Macoma* sp., *Mya* sp., and *Cryptobranchia* sp.

The assemblage of benthic foraminifers of the bed shows preponderance of *Elphidiella oregonensis* forms and indicates a younger (as compared to Bed I), probably early Pliocene age, of the host rocks.

Palynological data (Fig. 4) demonstrate that this interval of the section corresponds to the peak abundance of very diverse pollen of subtropical and Turgai floral elements. In this interval, we distinguished 13 exotic species missing in the present coastal flora. The pollen assemblage is rich in Juglandaceae (genera *Pterocarya, Platycarya, Carya, Engelhardtia,* and *Juglans*). Its other components are *Quercus* (two–three old species), *Ulmus, Tilia, Acer, Corylus*, and rare pollen grains of subtropical plants, such as *Castanea, Zelkova, Celtis, Rhus, Nyssa, Ilex, Aralia,* and Magnoliaceae, suggesting the ancient character of the assemblage. Moreover, the bed yielded at least three hornbeam species, including *Carpinus betulus* L., and also beech pollen representing *Fagus orientalis* Lipsky, *F. japonica* Maxim., and *F.* sp., all characteristic of humid monsoon zones. These deciduous broad-leaved taxa coexisted with diverse old species of angiosperms, and such a combination implies a subtropical type of climate.

We assume the early Pliocene age of the bed, because many genera of its flora are identical to the respective subtropical and Turgai taxa revealed by palynological analyses of coeval deposits in Primor'e (Sedova, 1957; Bolotnikova, 1979).

For instance, the late Miocene–early Pliocene flora from the Kidon Formation (Bolotnikova, 1979) is a possible age analogue of our assemblage, since both have approximately equal proportions of Pinaceae pollen. On the other hand, a similar flora of the southern Primor'e is, however, richer in Betulaceae, Juglandaceae, and *Alnus* species. In the opinion of Sedova (1957), the last feature was typical of the terminal warming episode of the Miocene, and her conclusion allows us to assume the early Pliocene age of the bed (5.1–4.3 Ma).

The revealed spore–pollen assemblage characterizes the early Pliocene climatic optimum with polydominant broad-leaved forests of the Turgai type, which accommodated some subtropical plants. The climate of that time was temperate to warm or monsoon-subtropical. Vegetation at the initial and terminal phases of this warming was represented by subtropical coniferous–broad-leaved forests that reflected the progradation of the higher cooler zones with the mountain taiga communities.

The D-alloisoleucine/L-isoleucine ratios in the *Mya* shells of the bed vary from 1.18 to 1.24. These values suggest a younger age of mollusks, as compared to those from Bed I.

At the time of the climatic optimum indicated by palynological data, the sea level was presumably higher (by 10–15 m) than at present, as is evident from the neritic type of sediments and from the character of buried fossils. It is conceivable that Bed II is coeval with the coastal terrace, which is 20–22 m high and composed of weathered pebblestones and compact sands.

Bed III (Figs. 2, 3). This bed of dense sands enclosing silt interlayers and shell detritus is stratigraphically discordant and 2 m thick. Its heavy mineral fraction insignificantly differs from that in the underlying sediments. Diatoms of the bed are not abundant and poorly preserved. Shells are highly fragmented and unsuitable for classification at the species level. Foraminifers are similar in composition to their assemblage in Bed II. Spore–pollen spectra of the bed show a higher diversity of small-leaved taxa and the appearance of cold-resistant components, such as certain Betulaceae, and thus suggest a much cooler climate than before. This cooling event separates two larger stages of warming in the early Pliocene. We failed to detect deposits like these in other drill sites.

Bed IV (Figs. 2, 3). The previous bed with cryophilic spore–pollen spectra is overlain by a rather thick (up to 6 m) sequence of dense sands with pebbles and abundant shells. The sand beds grade seaward into compact silt also bearing rock fragments and molluscan shells.

The heavy mineral fraction of these sediments is variable in composition and mainly consists of Fe–Mn-oxides and hydroxides. Near the base of the bed, it is enriched in rock detritus, sericite, and amorphous silica (up to 71%). Sedimentation proceeded under the active, but variable, conditions of water dynamics, and apparent stages of intense inflow of clastic material alternated with breaks in its deposition, as follows from the mineral analysis of the clay fractions.

The diatom assemblage of Bed IV is distinct because of the disappearance of *Cosmiodiscus insignis* and *Neodenticula koizumi*. The last occurrence levels of these species in the region is dated back to 4.4 and 3.7–3.5 Ma, respectively. The more thermophilic aspect of the assemblage is illustrated by certain species untypical of the arctic–boreal provinces, such as *Azpeitia tabularis* (Grun.) Fryxell et Sims, *Thalassiosira lineata* Jouse, *Actinocyclus ehrenbergii* Ralfs, *Rossiella tatsunokushiensis* (Koizumi) Gersonde et Shrader, and *Rhizosolenia styliformis* Bright, all newly appeared at this level. The assemblage also has a high percentage of species of the *Delphineis–Rhaphoneis* group, presumably indicating active water dynamics and a lesser depth than during the accumulation period of Bed II. *Coscinodiscus gorbunovii* Sheshuk., the freshwater species widespread in the Pliocene deposits of Primor'e, is another noticeable component of the diatom assemblage.

The diverse molluscan fossils of the bed are mainly represented by shallow-water forms rarely dwelling on bottom areas located deeper than 20–30 m. Their dominant species are *Mya* cf. *pseudoarenaria*, *M.* cf. *priapus*, and *Mactra* cf. *chinensis*. Both *Mya* species are very similar to their counterparts from the upper part of the Maruyama Formation in the coastal zone of Sakhalin. In addition, the molluscan assemblage includes other Neogene forms, such as *Fortipecten* sp., *Chlamys* cf. *cosibensis*, and *Anadra* sp. The general composition of the assemblage is characteristic of the early Pliocene and thus supports the age determination based on the diatom flora. A similar association of diatoms and malacofauna was reported from the lower Pliocene Tatsunokushi Formation of Honshu (Koizumi, 1973). The thermophilic character of both groups of fossils correlates with the early Pliocene climatic optimum recorded in the respective deposits of the Far East and North America (*Detal'noe* ..., 1992).

Spore–pollen spectra of the bed show preponderance of pollen of the diverse broad-leaved vegetation. In addition to old Pinaceae, we detected here at least eight main taxa: *Carpinus betulus* L., *Platycarya, Fagus, Engelhardtia, Carya, Shizandra, Tsuga*, and *Taxodium*. They indicate that the past coastal zone was covered with birch–broad-leaved forests of the Turgai type with an admixture of subtropical plants. Climatic conditions were humid and warm as in the present zones of subtropical biota.

The age of the deposits in question is well substantiated by both the palynological data and molluscan fauna. Two stages of warming distinguished in this work (Beds II and IV) may represent a single early Pliocene epoch of warming with an intermediate short episode of cooling (Bed III). We assume Bed IV to be as old as 4.3–3.7 Ma and correlative with an insignificant sea regression event.

Bed V (Figs. 2, 3). The bed is up to 7 m thick and composed of compact pelite and silt bearing rare pebbles and poorly preserved shells. Its section is most complete in drill Site 2. Authigenic minerals of the bed suggest a change of climatic conditions from the cooler initial to the warmer terminal phase. The water dynamics during the accumulation of sediments also changed from an active to a more calm state and was dependent on the sea-level fluctuations. This is recorded in the graded bedding of minerals according to their grain size and specific weight. In the lower part of the bed with oxidation conditions, clay minerals are less weathered.

Age determinations for the bed are based on the coexistence of *Neodenticula koizumi* and *Rossiella tatsunokushiensis* and on the parallel absence of some other diatom species characteristic of the second half of the Pliocene that originated approximately 2.5–3 Ma ago. In the middle latitudes of the Pacific, the last occurrence level of *Rossiella tatsunokushiensis* is dated back to 2.25 Ma, and *Neodenticula koizumi* probably appeared no later than 3.5 Ma ago. Using these data, we referred Bed V to the early–late Pliocene transition. The diatom assemblage is not homogeneous, and its thermophilic species are concentrated at two levels (15.5 and 12 m) in the lower and upper parts of the bed. These levels are separated by an interval where the thermophilic diatoms are missing.

At Site 2, the molluscan assemblage of the bed is almost monotaxonic. Mollusks are represented by *Mya* species, predominantly by *M.* cf. *pseudoarenaria* and *M.* cf. *priapus*. Some samples also yielded fragmented and seemingly redeposited shells of *Fortipecten* sp. The lack of real thermophilic forms in drill Site 2 along with abundant Balanidae and boreal *Liocyma fluctuosa* recovered by drilling at Site 1 are indicators of a minor regressive stage and climatic cooling.

Spore–pollen spectra of this level are enriched in boreal and periglacial components, such as the fruticulose *Betula* species, *Alnus, Larix,* and other taxa. Accordingly, they show that, in the coastal zone, the larch–alder forests substituted the birch–broad-leaved vegetation in response to the advancing cooling.

Thus, the results of our study allow us to refer the bed to the lower Pliocene age interval of 3.4–3.0 Ma.

Bed VI (Figs. 2, 3). This 4-m-thick bed is composed of sand and coarse-grained silt layers with pebbles and highly weathered shells. Authigenic minerals of the heavy fraction are represented here by oxidized Fe–Mn-compounds often composing pseudomorphs after pyrite crystals and shells. These minerals imply

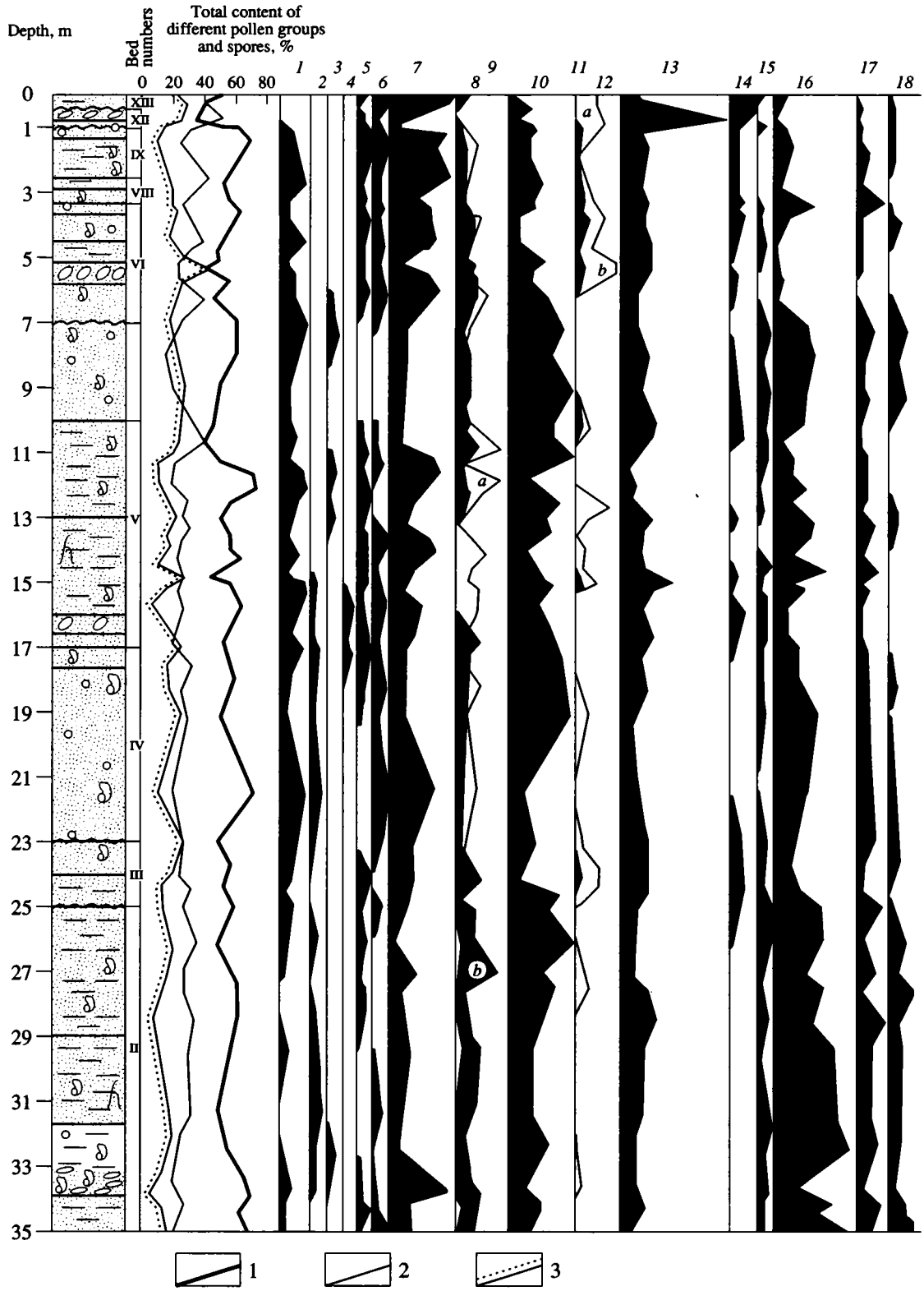
Depth, m
Bed numbers
Total content of different pollen groups and spores, %
0 20 40 60 80
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18
0 1 3 5 7 9 11 13 15 17 19 21 23 25 27 29 31 33 35
XIII XII IX VIII VI V IV III II
a b a b
1 2 3

Fig. 4. Spore–pollen diagram for Site 2 (plotted by G.M. Shumova): (1) pollen of arboreal and fruticose plants; (2) pollen of herbaceous plants; (3) spores; *1, Tsuga; 2, Taxodiaceae; 3, Cupressaceae; 4, Gliptostrobus; 5, Larix; 6, Abies; 7, Piceae; 8, 9, Pinus* s.g. *Haploxylon* (*a*) et s.g. *Diploxylon* (*b*); *10, Betula; 11, 12, Betula* sect. Fruticosae (*a*) et Nanae (*b*); *13, Alnus; 14, Alnaster; 15, Myrica; 16,* total abundance of broad-leaved pollen; *17, Quercus; 18, Ulmus; 19, Corylus; 20, Tilia; 21, Carpinus; 22, Fagus; 23 Juglans; 24, Pterocarya; 25, Platycarya; 26, Engelhardtia; 27, Zelkova; 28, Celtis; 29, Rhus; 30, Ostrya; 31, Carya; 32, Nyssa; 33, Castanea; 34, Shizandra; 35,* Araliaceae; *36, Ilex; 37,* Cyperaceae; *38,* Ericaceae; *39,* Graminae; *40,* Chenopodiaceae; *41, Artemisia; 42, Varia; 43, Bryales; 44,* Polypodiaceae; *45, Sphagnum; 46, Lycopodium; 47, Osmunda.*

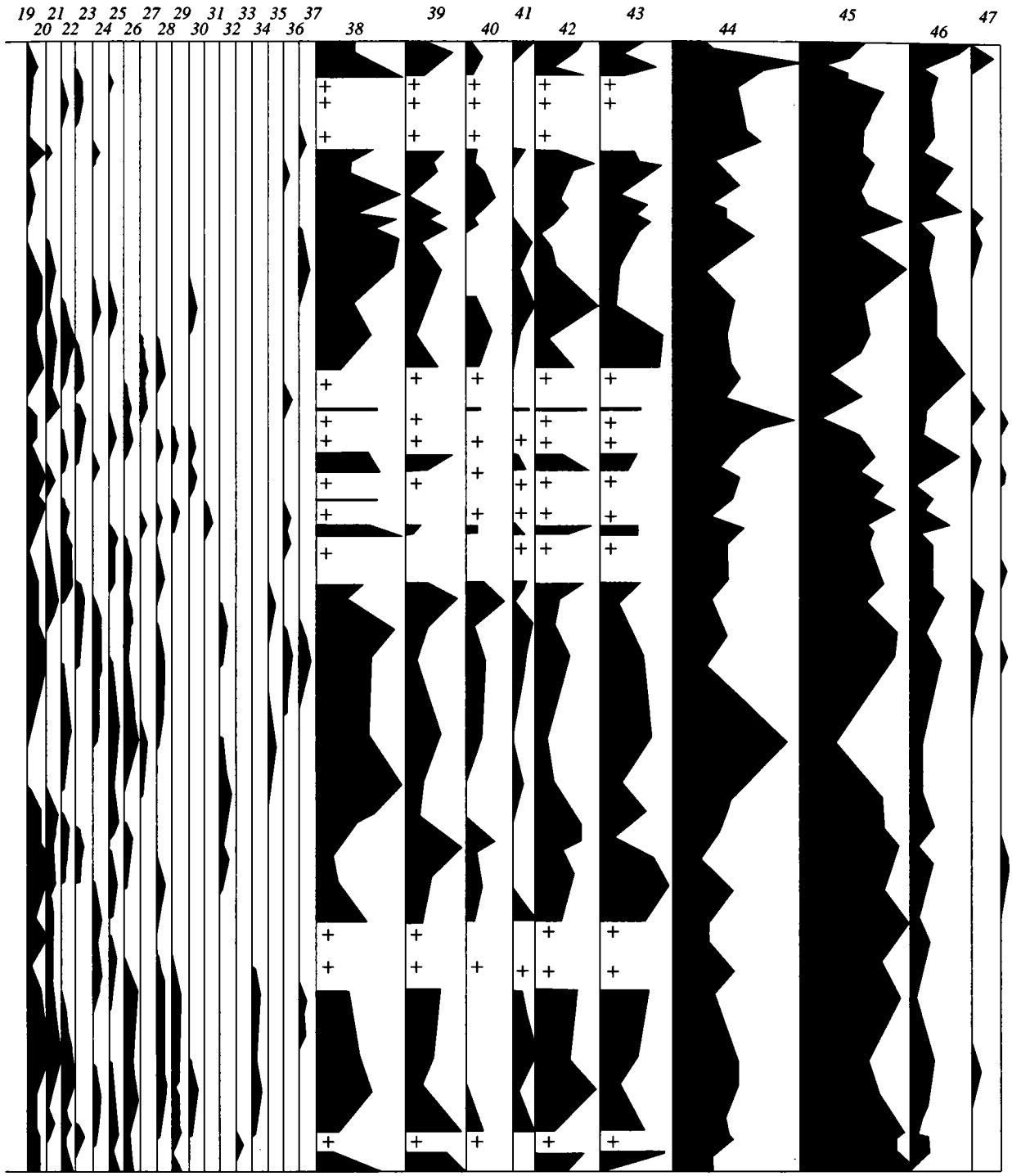

Fig. 4. (Contd.)

active water dynamics and oxidizing conditions of sedimentation (the transgressive stage).

The scarcity of diatoms restricts the correctness of age determination. The only indicative species is *Neodenticula koizumi,* whose stratigraphic range lies inside the Pliocene.

The molluscan assemblage is of a low taxonomic diversity. It includes *Mya* shells and redeposited detritus of extinct taxa, pointing to their burial in the upper subtidal zone.

Pollen assemblages at this level are more rich in arboreal taxa, most of all in diverse species of broad-

leaved plants. They suggest a wide distribution of the diverse birch–broad-leaved vegetation with some Neogene relics along the coastal zone. We detected only six taxa of exotic plants: *Fagus, Zelkova, Pterocarya, Rhus, Carpinus betulus* L., and *Tsuga*. The type of vegetation implies a new stage of considerable warming of the terminal Pliocene epoch. The erosion surface of the bed seemingly marks the Pliocene–Eopleistocene boundary.

Bed VII (Figs. 2, 3). This bed of layered sands with rare pebbles and abundant shells is up to 3.5 m thick. The composition of its heavy mineral fraction suggests rather calm reduction–oxidation conditions of sedimentation, probably reflecting a sea-level fall. By their composition, clay and authigenic minerals are indicators of a rather cool climate.

The upper part of the bed (beginning from the depth of 5 m in drill Site 2) yielded a diatom assemblage with approximately equal proportions of recent cryo- and thermophilic species, as is also typical of the region at present. The age of the assemblage corresponds to the stratigraphic range of silicoflagellate *Mesocena elliptica*. According to available data (Burckle, 1979), the datum level of this species is recorded everywhere in the central Pacific and spans the interval from the base of the Jaramillo Polarity Subzone to the base of the Brunhes Polarity Zone. In the geochronological scale, this corresponds to the interval of 1.3–0.79 Ma. The same datum level was detected by Saito and Burckle (1977) in the Wakimoto Formation of the Oga Peninsula (western Honshu).

Mya shells are again dominant here among mollusks. However, the depth interval of 4.0–4.6 m is very rich (95%) in fragmented thin-walled shells of the boreal species *Liocyma fluctuosa*. Taking into account the persistent lithology of sediments, we assume that this horizon corresponds to an episode of temperature decrease in the bottom water.

The general cooling of the climate is distinctly reflected in pollen assemblages of the bed. Their main components are species of small-leaved *Betula* and *Alnus*, especially of *Betula* sect. Fruticosae (up to 28%). At the same time, we recorded here the peak abundance of some periglacial taxa, such as *Betula* sect. Nanae (up to 8%) and *Alnaster* (up to 6%). The club-moss pollen (*Lycopodium pungens* L.) is confined to the same level. The associated pollen of conifers is poorly preserved, which allows us to consider it as derived from the Picea–*Pinus*–*Tsuga* forests populating mountain areas of the Sikhote Alin Ridge, i.e., far away from the coastal zone. The assemblage is of a rather cryophilic type, in contrast to past Pliocene pollen spectra. It seems to imply a considerable cooling in the Eopleistocene time.

Bed VIII (Figs. 2, 3). The bed is thin (1 m) and consists of brown silt enclosing pebbles and shells. The authigenic siderite preponderant in the heavy mineral fraction was definitely formed under oxidation conditions in a coastal lagoon with a quiet sedimentation environment.

The diatom assemblage of the bed is identical to that from Bed VII. The molluscan assemblage with dominant *Mya* species (mainly *M.* cf. *pseudoarenaria*) also includes shell detritus of *Cyclocardia crebricostata, Pododesmus macroshisma, Turritella* cf. *fortilirata, Neptunea* sp., and *Crepidula* sp. and redeposited fragments of thick-shelled *Fortipecten* sp. This assemblage represents the characteristic molluscan community of the middle and upper subtidal zones.

The spore–pollen spectra correspond to the warming stage that replaced the epoch of considerable cooling (Bed VII). They are again rich in arboreal pollen, mainly in that of *Betula* sect. Albae et sect. Costatae (up to 18%). *Quercus* is the dominant genus among broad-leaved taxa (16–26% in total). The additional components are *Betula* sect. Fruticosae (up to 17%) and *Alnaster* (up to 4%). Pollen of conifers is poorly preserved.

Vegetation in the provenance obviously became thicker at this time and was represented by broad-leaved, mainly birch–oak, forests on plains, by coniferous forests in mountains, and by fruticose tundra in the coastal zone. The respective warming stage corresponds to the Jaramillo Polarity Subzone and is dated back to 1.3–0.79 Ma on the basis of the diatom assemblage and associated silicoflagellate *Mesocena elliptica*.

Bed IX (Figs. 2, 3). This is the 2-m-thick bed of loose silts enclosing abundant and large fragments of shells. Its heavy mineral fraction includes up to 41% authigenic siderite, implying sedimentation under oxidation conditions in the calm closed lagoon.

Unfortunately, we do not have data on diatoms from this bed; however, it seems to be no older than the early Pleistocene according to its position above deposits spanning the magnetostratigraphic interval from the base of the Jaramillo Polarity Subzone to the base of the Brunhes Polarity Zone.

The molluscan assemblage of the bed is almost identical to that from Bed VIII.

In contrast, its palynological spectra are richer in pollen of dark conifers, mainly of *Picea* (26–36%), *Tsuga* (up to 16%), and *Larix* (up to 10%).

These features suggest preponderance of the dark fir–spruce taiga with accessory *Tsuga* species and broad-leaved plants in the vegetation cover of the provenance. It is likely that the coastal zone was occupied by birch–larch forests. Climatic conditions were moderately warm, and this stage can be correlated with one of the warming periods of the early–middle Pleistocene.

Bed X (Figs. 2, 3). This 10.5-m-thick bed is composed of silt with pellets and interlayers of sand. Its authigenic minerals are diverse. The lower part of the unit is rich in sulfides (up to 56%), which is characteristic of sediment accumulation under reducing conditions in water enriched in dispersed organic matter (a basin of the cold-water type). The main authigenic

component in the upper portion is siderite (up to 20%) formed in a lagoonal setting, where conditions were also reducing, but the water temperature was higher. Sediments of the bed also contain single grains of glauconite.

Unfortunately, no diatoms were observed in the bed (Site 6). Its lower horizon of sand incorporates redeposited fragments of the extinct Neogene mollusk *Fortipecten* sp. At the same depth level of 11.5 m, there are also single pollen grains of cryophilic and broad-leaved plants associated with the highly altered redeposited pollen. The latter feature and data of mineral analysis show that the rewashed components of older sediments are important ingredients of the sediments in question. In the uppermost part of the bed recovered at Site 6, we detected acute-angled fragments of *Mya* sp., *Macoma* sp., *Callithaca adamsi*, and some other shells.

We also distinguished six pollen zones in the main portion of Bed X using samples from Site 6. These zones correspond to particular phases of plant evolution, and when their pollen spectra are of the cryophilic type, they display approximately equal proportions of herbaceous, arboreal, and sporgenous taxa. In this case, the most abundant arboreal pollen is that of *Betula* sect Fruticosae (28–55%). *Alnaster* (8–57%) and the less abundant *Betula* sect. Nanae (up to 12%) and *Larix* (up to 15%) are preponderant representatives of the periglacial flora, whereas the sporgenous group is richest in *Sphagnum* and *Bryales* spores, and Cyperaceae are principal specimens of herbaceous plants.

During the respective phases, the coastal vegetation basically consisted of fruticose birch and sedge–moss tundra communities. It also included alder thicket and open larch-wood areas. These climatic phases were somewhat cooler and more humid than at present, as follows from the wide distribution of open coniferous forests with Dahurian larch in the current marsh areas (*Sovetskii* ..., 1971).

The relatively thermophilic palynospectra of the bed are considerably less rich in cryophilic taxa and include pollen of *Betula* sect. Albae et sect. Costatae (26–42%), *Alnus* (18–30%), and *Picea* (up to 12%) as the main components. The dominant Cyperaceae and rather abundant forbs (mainly Ericaceae) represent herbaceous plants of this flora. Moss spores are again its important constituents; however, Polypodiaceae (up to 28%) exemplify another significant group of the spore fraction from this level.

Accordingly, the vegetation cover of this type corresponded to that of the birch forest–tundra zone with alder–spruce communities concentrated in river valleys near the sea. The sedge–moss palustrine assemblages with additional Ericaceae were characteristic of the coastal zone. Climatic conditions were warmer.

Provided that broad-leaved plants and Neogene relics are missing in the sediments under consideration, we assume the Pleistocene age of the bed on the basis of the predominant pollen of small-leaved taxa associated with some periglacial components. On the other hand, the latter are not abundant, especially pollen of *Betula* sect. Nanae, and this, along with the nearly complete absence of herbaceous xerophytes, allows us to conclude that climatic conditions of this period were not as cold as, for instance, during the second cooling of the late Pleistocene.

We assume the middle Pleistocene age of Bed X (cold and intermediate climatic phases together). It is likely that the coastal terrace as high as 15–17 m, which is situated southward of the Syurkum Cape and mainly consists of large-pebbled material, was formed during this stage of shelf sedimentation.

Bed XI (Figs. 2, 3). This bed is conditionally distinguished in the coastal terrace of the Syurkum Cape, where the latter is 5–7 m high. Unfortunately, sediments of this geomorphological structure do not contain any fossils. In this respect, the terrace resembles other terraces of our region and Primor'e in general. They usually contain very poor paleontological material, and their genesis is often studied by means of lithologic analysis. In our case, we used the morphoscopy of quartz grains. The results obtained show that, judging from the grain shapes, their roundness, and the character of the surface, and also from the weathering degree and structural parameters of the sediments, the lower pebblestone layer of the terrace is of marine origin, whereas its upper loam layer with boulders has distinct features of diluvial–proluvial sediments.

Our age evaluation of the Syurkum Cape terrace is based on the results of TL dating. The age values obtained are 128000 ± 32000 (TL-557) and 109000 ± 27000 (TL-558) years for the lower pebble and upper loam layers, respectively. These values correspond well to known data on the maximum sea level rise by 10–15 m above the present level during the Kazantsevo transgression in the Sea of Japan.

Accordingly, we accept the late Pleistocene age of the terrace as correlative with the Kazantsevo interglaciation and respective intense transgression.

Bed XII (Figs. 2, 3). Sediments of the bed (up to 1 m thick) are represented by sand enclosing pebbles and shell remains. The heavy mineral fraction is almost lacking authigenic minerals (1% only). Its terrigenous components are diverse in composition and well to poorly rounded, and fragments of epidote rocks occurring among them are oxidized. The terrigenous material is well sorted and suggests an active sedimentation hydrodynamics.

The diatom assemblage of the bed does not include extinct species, except single specimens of the redeposited Miocene forms. *Paralia sulcata* dominant among diatoms is typical of the present-day bottom surface at Site 2. On the other hand, the obvious association of redeposited Miocene–Pliocene taxa with currently dominant pelagic species can be interpreted as having formed during the coldest epoch of the late Pleistocene, i.e., during the Sartan cooling.

The abundant shell detritus is represented at this level by disintegrated *Mya* shells, among which we detected one fragment of the Pliocene form *Thyasira* sp.

Spores and arboreal pollen are equally abundant in palynospectra of the bed. The spectra correspond to the vegetation with widespread alder thickets and open larch–birch forests of the cold and humid climate. Palynological data may indicate a cooling event of the late Pleistocene time, probably the Sartan event, as suggested by diatom analysis. The radiocarbon dating of molluscan shells from the bed yielded age values of 34 400 ± 1200 (GIN-7244), 34 100 ± 1000 (GIN-7243), and 11 380 + 170 (MSU-1377) years.

Bed XIII (Figs. 2, 3). This is a 0.5-m-thick bed of silt with fragmented shells. Its heavy mineral fraction is identical in composition to that of Bed XII. The diatom assemblage corresponds to respective flora existing near the bottom under the current thermal and salinity conditions of the basin. The shell detritus includes fragments of contemporaneous species *Callithaca adamsi, Patinopecten yessoensis, Macoma* sp., *Mya* sp., *Serripes* sp., and *Cryptonatica* sp. Pollen and spores correspond to derivatives of the dark coniferous taiga of the fir–spruce type. The inferable climate and vegetation correspond to the current environment, and we refer the bed to the uppermost Holocene.

CONCLUSION

The results obtained allow us to conclude the following:

(1) We established that two-thirds of the composite section of the shelf sedimentary cover near the Mednyi Cape of the western Tartar Strait is composed of thick marine Miocene–Pliocene deposits.

(2) The age interval of this sedimentary sequence is substantiated by data on the distribution of diatoms, spores, pollen, and malacofauna correlated with other geological data.

(3) The sequence of Neogene sediments seems to be preserved in a narrow depression having the subsidence amplitude of 0–500 m and extending along the coast from Sovetskaya Gavan' to Nikolaevsk-na-Amure (*Yug dal'nego* ..., 1972).

(4) The correlation of deposits, which were recovered by drilling at Sites 1 and 2, showed that the entire sequence of Pliocene–Eopleistocene sediments experienced dislocations in the course of the early Pleistocene folding and block movements in the region.

(5) The terminal late Miocene (older than 5 Ma) corresponded to the period of climatic cooling. Drilled sections include sediments deposited at the final stage of this period, when climatic oscillations (alternation of cooler and warmer phases) provoked the respective changes in vegetation. The general sea-level rise at this time (presumably the beginning of the early Pliocene transgression) was somewhat fluctuating, as follows from the composition of molluscan fauna and from the structure of the transgressive sequence.

(6) The peak of the early Pliocene transgression occurred at the time of the climatic optimum revealed by palynological data. It was the accumulation time of relatively deep-water deposits (the inferred basin depth was more than 100 m) with the neritic diatom assemblage. It is also probable that the coastal marine terrace 20–22 m high was formed at the same time.

(7) The second early Pliocene warming was separated from the first one by a cooling event. It was less significant and developed at the stage of sea regression. The respective sand bed includes thermophilic diatom and molluscan assemblages; the latter inhabited the upper subtidal zone. The interval of beds II–IV most likely corresponds to the complete climatic optimum with an intermediate cooling episode. This was the third optimum of global character during the Neogene period. It is marked by the appearance of southern boreal diatoms and palynological assemblages with abundant thermophilic components. At this time (about 4 Ma ago), climatic and paleogeographic belts were displaced northward by 10°–12°, and the Bering Strait was probably flooded due to the high sea level (*Detal'noe* ..., 1992).

(8) The subsequent history of sea-level oscillations is less evident. We assume a regression stage at the beginning of the late Pliocene (bed of respective deposits) followed by an insignificant sea-level rise during the terminal late Pliocene episode of warming.

(9) During the Olduvai cooling of the Eopleistocene, there was a regression stage (respective deposits with dominant arctic–boreal *Liocyma fluctuosa* and specific authigenic minerals) that is recorded in the "most cryophilic" spore–pollen spectra. The warm Jaramillo epoch was responsible for lagoon formation in the region.

(10) Lagoons were also formed during the cold epochs of the early–middle Pleistocene.

(11) The available data on deposits of the last interglaciation (basal upper Pleistocene) are insufficient for paleogeographic inferences. We assume, however, that this Kazantsevo time was responsible for the formation of the marine terrace 5–7 m high at the Syurkum Cape.

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The Late Stage of Evolution of the Northern Part of the Okhotsk–Chukchi Volcanogenic Belt

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Abstract—As a result of field studies conducted in the Enmyvaam River basin in 1991 and 1993, it has been found that stratigraphic units of volcanic rocks of the late stage of the Okhotsk–Chukchi volcanogenic belt (OCVB) evolution make up a succession (from the base upward) of the Ergyvaam, Emuneret, and Enmyvaam formations. The work also resulted in obtaining new palynologic and magnetostratigraphic data. Rocks with reversed magnetization (the middle part of the Emuneret Formation) were revealed and observed for the first time among volcanic complexes of the OCVB. The rocks occur stratigraphically above layers with the Ust'-Emuneret floral assemblage referred by palynological data to the late Turonian. The reversed polarity zone (R-zone) revealed correlates with the Klyuev R-zone (the middle Coniacian) of the general paleomagnetic scale for the Phanerozoic of continents. The lower part of the Enmyvaam Formation comprises the late Turonian–early Senonian palynological assemblage. Taking into account the stratigraphic position of the Enmyvaam Formation and the normal magnetization of its rocks, the formation is dated back to the Santonian. This was the first attempt to apply the combined paleophytological, palynological, and magnetostratigraphic data for correlation the Cretaceous volcanogenic deposits in northeastern Russia.

Key words: *Okhotsk–Chukchi volcanogenic belt, late stage, Turonian, Coniacian, Santonian, floral assemblage, palynological assemblage, magnetostratigraphy, volcanology.*

INTRODUCTION

The northern part of the OCVB is distinguished in this work as the Central Chukchi sector (CCS), in which the external and internal zones are distinguished like in many other parts of the OCVB (Fig. 1).

The external zone comprises strata corresponding only to the early stage of evolution of the CCS. The strata make up the Chaun Group with abundant fossil flora, which is referred to as the Chaun floral assemblage (Belyi, 1977). The flora is dated back to the late Albian (Belyi, 1977; Samylina, 1988).

In the internal zone, the Chaun Group is overlain, with a hiatus and structural unconformity, by volcanic rocks of the late evolutionary stage. Belyi and Raikevich studied in detail these rocks in escarpments of the Enmyvaam River in 1991 and 1993. The palynological analysis of the selected samples was made by B.V. Belaya.

STRATIGRAPHY

According to the Resolutions of the 2nd Interdepartmental Regional Conference on Stratigraphy (1974–1975), it was agreed to subdivide the volcanogenic deposits developed in the middle course of the Enmyvaam River into three parts (from the base upward): the Emuneret (Snezhnaya) sequence, and the Ergyvaam and Enmyvaam formations (*Resheniya* ..., 1978). In more recent times of the progress in medium-scale geologic mapping, the Emuneret sequence was uprated to the rank of formation. However, the relationships of these stratigraphic units remained unstudied until recently.

As a result of field studies, the Emuneret Formation was found to occur between the Ergyvaam and Enmyvaam formations. Thus, the stratigraphic units of volcanic deposits of the late evolutionary stage of the CCS in the OCVB, which were distinguished earlier, make up a succession of the Ergyvaam, Emuneret, and Enmyvaam formations (Fig. 2).

The Ergyvaam Formation occurs above the Koekvun, or more rarely, Pykarvaam formations of the Chaun Group. In the vicinity of Elgygytgyn Lake and north of it, the Ergyvaam Formation is composed of closely alternating greenish tuffs; acidic ignimbrites and tuffstones (100–300 m thick, the lower part); and reddish, brownish, or sometimes gray rhyolite ignimbrites, among which black vitrophyres are found (200–300 m thick, the upper part). *Coniopteris* cf. *tschuktschorum* (Krysht.) Samyl., *Coniopteris* sp., *Asplenium* cf. *dicksonianum* Heer, *Cladophlebis frigida* (Heer) Sew., *Cladophlebis* cf. *grandis* Samyl., *Ginkgo* aff. *adiantoides* (Ung.) Heer, *Sphenobaiera* ex gr. *logifalia* (Pom.) Fl., *Phoenicopsis* ex gr. *angustifo-*

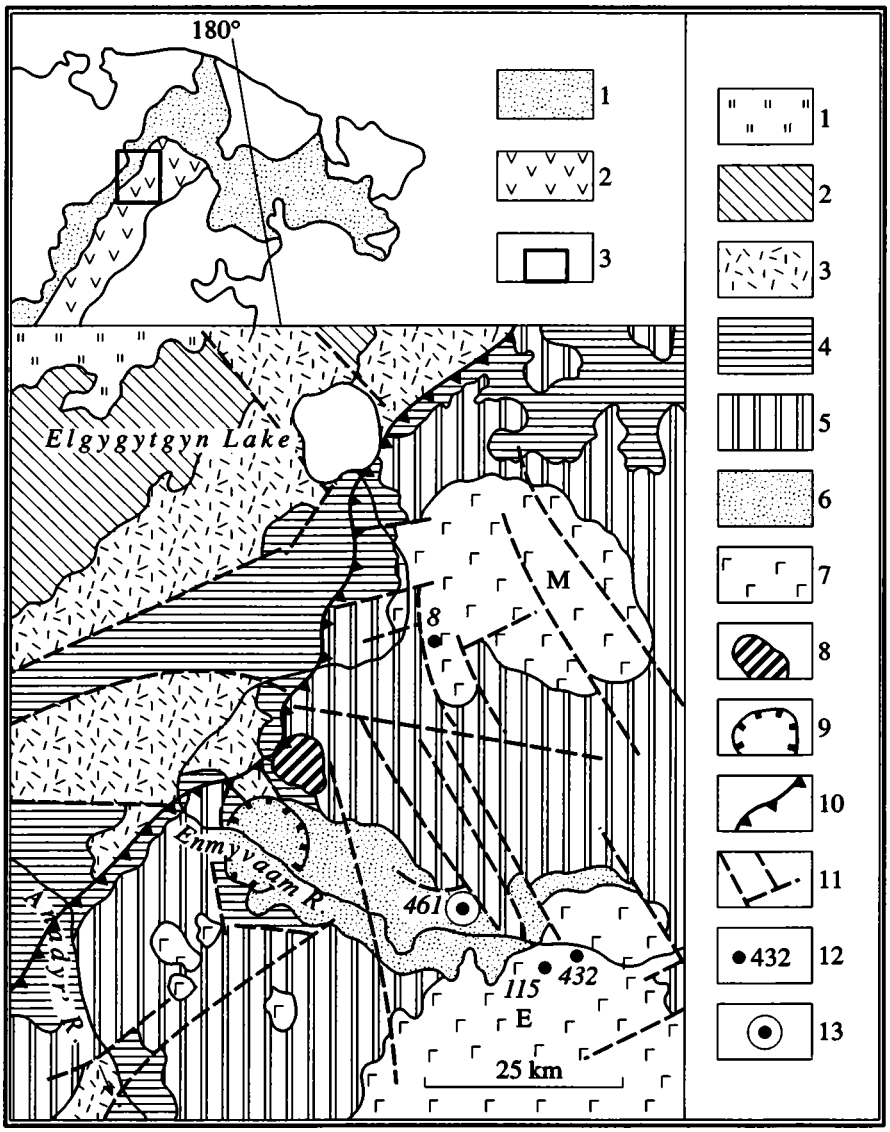

Fig. 1. The geologic structure of the southern part of the Central Chukchi sector of the OCVB. The inset map shows (1) external and flank zones; (2) the internal zone of the OCVB; (3) location of the region shown in the main figure. Symbols for the main figure: (1) the orogenic complex of Chukchi Mesozoides (the Pycheveem Formation); (2–7) volcanic rocks of the OCVB: (2–4) rocks of the early evolutionary stage (the Chaun Group) including (2) the Al'kakvun (Ugatkyn) and Kalen'muvaam formations, (3) the Pykarvaam and Voron'ya formations, and (4) the Koekvun Formation; (5–7) rocks of the late evolutionary stage including (5) the Ergyvaam Formation, (6) the Emuneret Formation, and (7) the Enmyvaam Formation; (8) granodiorite–porphyry; (9) the Snezhninskaya Caldera; (10) the boundary between the external and internal zones of the OCVB; (11) faults; (12) observation point and its number; (13) locality of the Ust'-Emuneret fossil flora; (M) the Mechekrynnetveem and (E) Enmyvaam fields of aluminiferous plateau basalts.

lia Heer, *Taxites intermedius* (Holl.) Samyl., *T.* ex gr. *intermedius* (Holl.) Samyl., *Torreya gracillima* (Holl.) Krysht. et Baik., *Sequoia ambigua* Heer, *S.* cf. *concinna* Heer, *Thuja cretacea* (Heer) Newb., and *Quereuxia angulata* (Newb.) Krysht. were collected from the member of volcanosedimentary rocks in the lower part of the formation exposed in the upper course of the Mil'guveem River, on its left bank, north of the region shown in Fig. 1 (Belyi, 1977). This taphoflora is similar to that present in the Ola Formation of acidic volcanites in the Magadan region of the OCVB and was assigned by Samylina (1988) to the Arkagala stratoflora (1988).

South of Elgygytgyn Lake, the Ergyvaam Formation consists predominantly of massive gray, brownish, and reddish porphyroclastic rhyolite ignimbrites, among which major bodies of black vitroignimbrites and rare nonpersistent tuff members occur; the thickness is 600–700 m.

Fig. 2. Geologic structure of the area in the middle course of the Enmyvaam River between the Snezhnaya Caldera and Basalt Cliffs (Point 432): (1–2) volcanic deposits of the early evolutionary stage of the OCVB including (1) the Pykarvaam Formation and (2) the Koekvun Formation; (3–11) volcanic deposits of the late evolutionary stage of the OCVB: (3, 4) the Ergyvaam Formation of (3) massive rhyolite ignimbrites and (4) ignimbrites, tuffs, and rhyolite vitrophytes; (5–8) the Emuneret Formation, including its (5) lower, (6) middle, and (7) upper parts described in the text, and also (8) undifferentiated deposits of this unit; (9) the caldera complex (the Snezhnaya Caldera, see description in the text); (10) the lower and (11) upper parts of the Enmyvaam Formation as described in the text; (12, 13) subvolcanic intrusions, such as (12*a*) stocks and (12*b*) basaltic dikes of the Enmyvaam Formation, and (13) moderately acid stocks of the Enmyvaam (13*a*) and Emuneret (13*b*) formations; (14) the tuff member at the base of the middle part of the Emuneret Formation; (15) boundaries of the Snezhnaya Caldera (faults); (16) rupture dislocations; (17) brecciation zone; (18) stratigraphic boundaries between pyroclastic flows; (19) observation point and its number; (461) point of the paleomagnetic test; (112) point of supplementary geologic observations; (20*a*) locality of the Emuneret palynological assemblage and the Ust'-Emuneret flora; (20*b*) locality of the Enmyvaam palynological assemblage (Basalt Cliffs); (20*c*) locality of the Merevaam palynological assemblage.

The Emuneret Formation is not as widespread as the Ergyvaam Formation. On the left bank of the Enmyvaam River, near the Emuneretveem River mouth (Fig. 2, Site 461), thinly alternating tuffstones and acidic tuffs of the Emuneret Formation occur without visible unconformity on porphyroclastic ignimbrites of the Ergyvaam Formation. A complete section of the Emuneret Formation has been compiled for the first time and is described as follows:

The Lower Part

Unit *em-1*. A member of gray tuffstones and acidic tuffs, 42 m thick.

Unit *em-2*. A member of green tuffs, tuffstones, tuff breccias of composite structure, white and light gray ash tuff and tuffstones, 300 m thick. The upper part of the member comprises different fossil plants collected by Lebedev and referred to as the Ust'-Emuneret assemblage (Lebedev, 1987) of *Arctopteris* aff. *rarytkensis* Vassil., *Cladophlebis* sp. 1–3, *Sphenopteris* sp. 1–2, *Ginkgo* cf. *adiantoides* (Ung.) Heer, *Ginkgo* sp., *Phoenicopsis* ex gr. *angustifolia* Heer, *Sphenobaiera* spp., *Araucariters* sp., *Cephalotaxopsis* cf. *magnifolia* Font., *Cephalotaxopsis* sp., *Cedrus* (?) sp., *Pityospermum* sp. 1-2, *Glyptostrobus* sp., *Sequoia* sp., *Libocedrus* (?) sp., *Trochodendroides arctica* (Heer) Berry, *Trochodendroides* spp., *"Quercus" tchucotica* Abram., *Macclintockia* sp., *Menispermites* sp., *Ziziphus* sp., *Platanacea* sp., *"Paliurus"* sp., *Dictyledones* spp., and *Monocotyledones* sp. By the presence of *"Quercus" tchucotica*, the Ust'-Emuneret assemblage is correlated with flora of the Barykov Formation and dated back to the early Campanian (Lebedev, 1987).

Unit *em-3*. A member of white, thin-platy, acidic ash tuffs, 35 m thick.

Unit *em-4*. Massive, crystalloclastic rhyolite ignimbrites, 30 m thick.

The Middle Part

Unit *em-5*. White ash tuffs replaced upsection by lapilli tuffs and acidic tuff-breccias, 60–70 m thick.

Unit *em-6*. Thin-bedded, acidic ash tuffs, 30 m thick.

Unit *em-7*. Black rhyolite vitroignimbrites, 12 m thick.

Unit *em-8*. Light lilac and violet, lenticular and banded rhyolite ignimbrites, 40 m thick.

The Upper Part

Unit *em-9*. Massive and banded rhyolite ignimbrites alternating with black vitroignimbrites; the upper part comprises white tuffs and rhyolite tuff breccias. The thickness is 60–70 m.

Unit *em-10*. A member of thin-bedded, brownish red tuffstones, tuff siltstones, basic and intermediate psammitic and lapilli tuffs, 35 m thick.

Unit *em-11*. Green and pinkish tuffs, tuffstones, ignimbrites and black acidic vitrophyres, 70 m thick.

The total thickness of the Emuneret Formation is 430–450 m.

The Snezhnaya Caldera mapped in the western part of the Emuneret Formation area (see Fig. 1) is considered as a source (most likely as one of the sources) of stratified volcanic deposits characterized above. The detailed description of the caldera will be published elsewhere. In what follows, only a brief description of principal features of the structure and composition of rocks will be given.

In the central part of the caldera, the lower visible unit of volcanic deposits is made up of white and yellow tuffs and acidic tuff breccias. They are similar to tuffs and tuff breccias, which the middle part of the Emuneret Formation begins with (see the section description, Unit *em-5*).

In the northern part of the caldera, the tuffs are overlain by fluidal rhyolite lavas and vitrophyres, which make up a major extrusive dome. In the southern part of the caldera, the massive crystalline rhyolite ignimbrites, which comprise numerous beds, dikes, and stocks of black vitroignimbrites, are exposed above the tuffs. Various unstratified and/or slightly stratified tuff breccias and tuffs and dikelike bodies of rhyolites, dacites, and vitroignimbrites are developed in the peripheral walls of the caldera. Zones of intense sulfidization, oxidation, and argillization are abundant in the inner part of the caldera, especially along its edges. All deposits infilling the caldera are intruded by dikes of olivine basalts.

The Enmyvaam Formation belongs to the aluminiferous plateau basalts, which commonly crown the sequence of volcanic deposits in the OCVB (Belyi, 1994). It usually overlies either ignimbrites of the Ergyvaam Formation or andesite basalts and tuffs of the Koekvun Formation. In the Enmyvaam River valley, basalts of the Enmyvaam Formation occur directly on deposits of Unit *em-11* of the Emuneret Formation. The Enmyvaam Formation section is as follows:

The Lower Part

Unit *en-1*. Flows of highly vesicular, amygdaloidal and massive olivine basalts, 130–150 m thick.

Occurring above this unit is a complicated rock sequence of contrast composition, which has been studied in escarpments of Basalt Cliffs (Figs. 1, 2, Site 432; Belyi, 1994, Fig. 8).

Unit *en-2*. A bed (~10 m thick) of moderately acidic ignimbrites occurs at the base, being overlain by flows of massive and amygdaloidal basalts about 100–110 m thick.

Unit *en-3*. Thin-bedded basaltic tuffs and tuff breccias (~35 m) grading upward into a member of thin-bedded tuffstones and siltstones (~55 m). Packets of thin-bedded acidic tuffs are observed among basaltic tuffs and sedimentary rocks. The thickness is 90 m. Abundant imprints of *Thallites* sp., *Sphenopteris* sp. cf. *Aspelenium dicksonianum* Heer, *Phoenicopsis* ex gr. *angustifolia* Heer, *Thuja cretacea* (Heer) Newb., and *Crataegites* sp. occur in tuffstones and siltstones 25 m below the top of the member. Despite the poor taxonomic composition, this locality of fossil plants can be attributed with a high probability to the Arkagala stratoflora, rather than to any other. Taphoflora representing a depleted modification of the Arkagala-type paleoflora was also found in plateau basalts of the Mygdykit Formation in the Magadan area of the OCVB (Belyi and Samylina, 1987).

Unit *en-4*. Basalts (12 m) overlapped by a packet of dark gray sticky clays with carbonized plant detritus and slightly cemented gravelstone (8 m); the thickness is 20 m.

Unit *en-5*. Massive, porphyroclastic, moderately acidic ignimbrites and xenoignimbrites with rare tuff interlayers; the thickness is 60 m.

Unit *en-6*. A member of acidic tuffs, tuffstones, tuff siltstones, and soaking mudstones, which comprise coal lenses; the thickness is 20 m.

Unit *en-7*. Pinkish, slightly welded, crystallo–vitroclastic tuffs and rhyolite ignimbrites, 22 m thick.

Unit *en-8*. Boulder and gravel volcaniclastic conglomerates (a lenslike body 10 m thick) overlain by a basalt flow (the visible thickness is 12 m). These deposits belong to the upper part of the Enmyvaam Formation, which was observed in the Merevaam River basin. The total thickness of the lower part of the Enmyvaam Formation is 450–470 m.

The upper part of the Enmyvaam Formation was studied in the basin of the Merevaam River, the right tributary of the Enmyvaam River (Fig. 1, Site 115). This is a uniform series of flows of massive and vesicular basalts with a visible thickness up to 500 m. The base of the sequence was not directly observed here. Among basalt flows of its lower part, there are lenticular bodies (4 m thick and 30 m long), which are composed of tuff siltstones and tuffstones with thin coal interlayers, and also with sedimentary volcaniclastic breccias and gritstones.

Further northward, in the basin of the Mechekrynnetveem River, a field of olivine basalts, which form a well expressed isometric negative volcanic structure of subsidence, belongs to the Enmyvaam Formation. The visible thickness of basalts is 150–200 m; they overlap ignimbrites of the Ergyvaam Formation. Only in the western part of the volcanic structure (Fig. 1, Site 8) do basalts occur on sedimentary rocks with a visible thickness of 50 m. The sedimentary rocks are represented by dark gray, easily soaking siltstones, sandstones, and mudstones, which comprise abundant plant detritus and imprints of *Quereuxia angulata* (Newb.) Krysht. White and yellow thin interlayers, probably of ash tuffs transformed into a clayey mass, are observed among the sediments. Siltstones are overlain by a layer (3–3.5 m) of bright yellow conglomerates composed solely of pebbles of ignimbrites and acidic lavas. A layer of dark gray tuff gritstone (0.5 m), made up of well-rounded gravel (acidic volcanites) and basaltic tephra, occurs between the conglomerates and basalts.

RESULTS OF PALYNOLOGICAL STUDIES

The study of sections of the Emuneret and Enmyvaam formations in the Enmyvaam River basin was accompanied by a regular sampling for palynological analysis. Six samples from the Emuneret Formation (4 samples from Unit *em-2* and 2 samples from Unit *em-10*), as well as 18 samples from the Enmyvaam Formation (10 samples from Unit *en-3*, 3 samples from Unit *en-4*, 2 samples from Unit *en-6*, and 3 samples from the upper subformation, Site 115), totalling 24 samples, were taken and prepared for analysis. Twelve samples appeared to be representative, with abundant and diverse microfossils, each sample comprising from 700–800 to 1600–1800 grains. They are shown in Fig. 3. Other samples, each containing less than 200 grains, were recognized as unrepresentative. Besides, 10 representative palynological samples were taken from siltstones and mudstones exposed in the Mechekrynnetveem River basin. Localities of representative palynological samples are shown in Figs. 1 and 2, Sites 461, 432, 115, and 8.

Three palynological assemblages are distinguished in the revealed spores and pollen spectra as corresponding to the Emuneret (Unit *em-2*), Enmyvaam (units *en-3*, *en-4*, and *en-6*), and Merevaam (the upper subformation of the Enmyvaam Formation, Site 115) palynofloras. Spectra of ten samples taken in the Mechekrynnetveem River basin are referred to the Enmyvaam palynological assemblage.

The Enmyvaam palynological assemblage is defined in deposits that are spaced apart, vary in composition, differ in structure, and have considerable thickness. The characteristic and indicative composition of the palynological assemblage offers a reliable identification of its analogues in the Cretaceous sections of adjacent territories. The age of deposits with the Emuneret and Merevaam palynological assemblages is determined mostly by their stratigraphic position with respect to the sequence comprising the Enmyvaam palynological assemblage. Thus, it makes sense to describe this palynological assemblage first.

The Enmyvaam palynological assemblage is distinguished by eight samples taken in Basalt Cliffs (Figs. 1, 2, Site 432) and ten samples selected from siltstones and mudstones in the Mechekrynnetveem River basin (Fig. 1, Site 8).

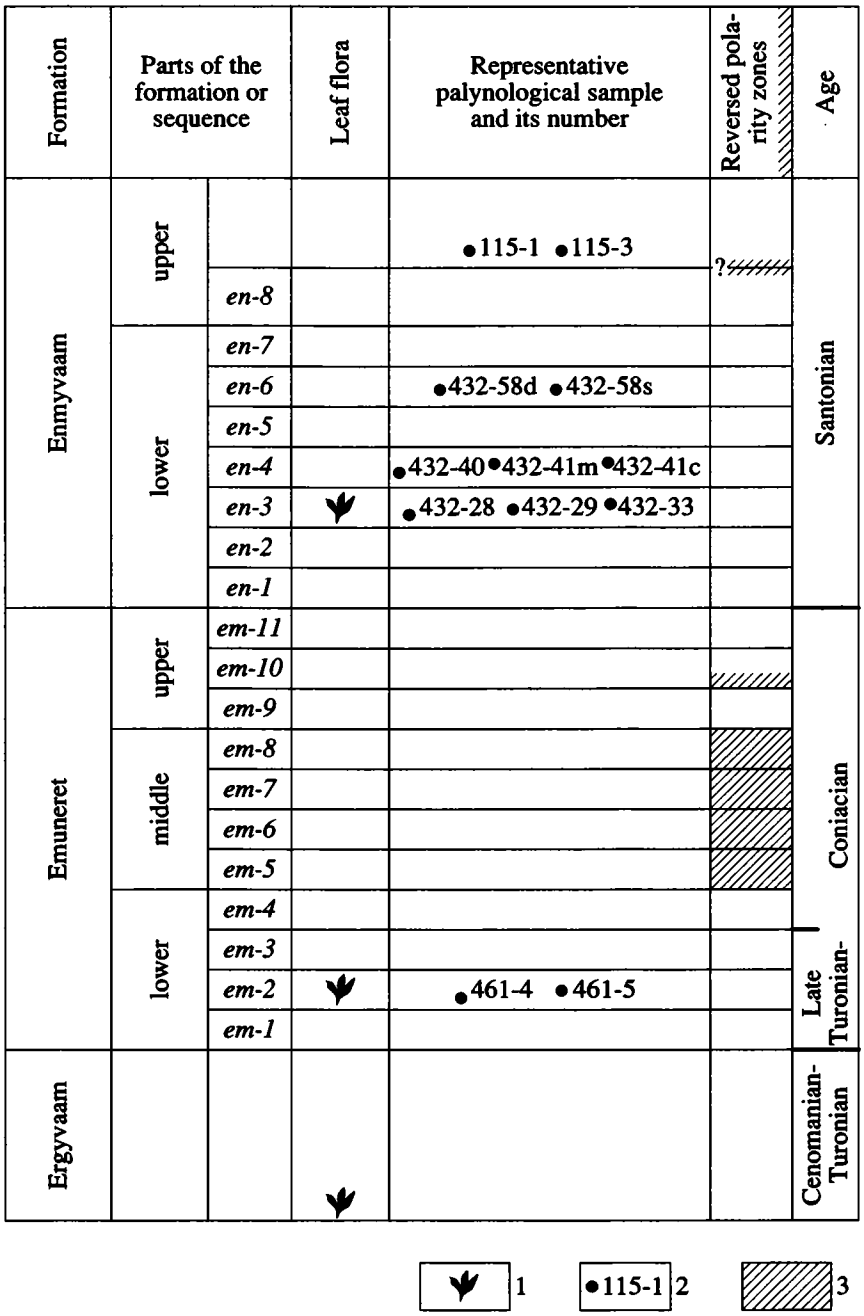

Fig. 3. Localities of leaf flora and palynological samples with respect to reversed polarity zones in the stratigraphic sequence of volcanic deposits of the late CCS evolutionary stage of the OCVB. (1) Leaf flora; (2) locality and number of palynological samples; (3) rocks with reversed magnetization.

In the Basalt Cliffs exposure, the samples were taken from stratigraphic units *en-3, en-4,* and *en-6* in the 140-m-thick interval (Belyi, 1994, Fig. 8); the thickness of the deposits from which the samples were taken in the Mechekrynnetveem River basin is about 50 m. The Basalt Cliffs section comprises packets of sticky clays and soaking siltstones similar to those developed in the Mechekrynnetveem River basin; palynological samples from these deposits are the richest in microfossils. The Enmyvaam palynological assemblage is characterized by the following:

(a) A high content of spores (15.5–96.6%, or 50.9% on average) of diverse and abundant transit taxa, mainly *Laevigatosporites*, associated with *Leiotriletes, Sphagnumsporites* and *Osmundacidites* and sometimes with *Retitriletes, Lycopodiumsporites,* and *Cyathidites.* Found in a smaller amount are (1) *Concavisporites junctus* (K.-M.) E. Sem., a relict in the Cretaceous

deposits; (2) *Cleichtniidites, Cicatricosisporites, Schizaea, Lygodiumsporites subsimplex* (Naum.) Bolch., and others, all common for the Lower and Upper Cretaceous deposits; (3) *Taurocusporites reduncus* (Bolch.) Stover, *Foveosporites senomanicus* (Chl.) Schwetz., *Selaginellidites spinulosus* (Cook et Dett.) var. *heberatus* Krasn., *Divisisporites euskirchenensis* Thomson, *Cingulatisporites euskirchensoides* Del. et Sprum., and others, all confined to the upper horizons of the Lower and the lower half of the Upper Cretaceous and found sometimes even in the Senonian;

(b) A high pollen content of gymnosperms (3.0–83.7%, or 47.9% in average) represented by diverse bisaccate taxa (*Pinuspollenites, Alisporites, Piceapollenites, Abiespollenites, Cedripites, Dacrydiumites, Phyllocladidites,* and *Podocarpidites*), Taxodiaceae, and undivided Taxodiaceae–Cupressaceae. *Podozamites* sp., *Ginkgocycadophytus* sp., *Araucariacites* sp., and other forms are rarely found;

(c) An insignificant pollen amount of angiosperms (0.4–10.5%, averaging 1.2%), among which are tricolpate and tricolporate taxa typical of the Cenomanian–Turonian deposits. Extremely rare are pollen grains of formal genera of the "unica" morphological type: *Parviprojectus dollium* Samoil., *Integricorpus bellum* N. Mtch., *Fibulapollis mirificus* Chl., *Aquilapollenites* sp., *Azonia fobacea* Samoil., and *Gothanipollis* sp., which appear for the first time in the upper Turonian–lower Senonian deposits. Forms resembling pollen of *Betula, Myrica, Alnus, Quercus, Acer,* Platanaceae, and Menispermaceae are very rare.

Similar palynological assemblages have been identified in flora-bearing deposits of the Arkagala Formation in the Tal-Yuryakh quarry (Markevich, 1989) and in deposits of the Pervomai Formation (Belyi and Samylina, 1987) underlying plateau basalts of Gipoteticheskii Creek. The Enmyvaam palynological assemblage is similar to the palynological assemblage from the Turonian–lower Senonian deposits of the lower and middle subformations of the Bykov Formation in Sakhalin (Bityutskaya *et al.,* 1973; Budrin, 1969), and also to palynoflora from the lower Senonian deposits of the Verkhnii Zavitinsk Formation in the Zeya–Bureya depression (Khlonova, 1969, Hole 2-k; Bratseva, 1969, Hole 21-k). It should be emphasized that the lower Senonian palynological assemblages of the Verkhnii Zavitinsk Formation are more diverse in composition and have a higher content (5–28%, Khlonova, 1969; 15–20%, Bratseva, 1969) of pollen of angiosperms, as compared to the Enmyvaam palynological assemblage. Therefore, the Enmyvaam assemblage, like the Arkagala flora, is dated back to the late Turonian–early Senonian.

The Emuneret palynological assemblage is distinguished by two samples, 461-4 and 461-5 (see Fig. 2, Site 461), taken from flora-bearing beds in trenches, where Lebedev collected the Ust'-Emuneret floral assemblage (Lebedev, 1987).

The pollen of gymnosperms dominates in the assemblage (73.9–79.4%), spores make up 20.0–24.5% here, and the pollen of angiosperms amounts to 0.6–1.6%. Among gymnosperms, the amount of pollen representing the *Taxodiaceaepollenites–Cupressaceaepollenites* group (33.5–47.5%) and that representing bisaccate conifers (31.6–40.3%) were found to be approximately equal. Forms dominating among conifers are species of genera *Pinuspollenites, Alisporites, Piceapollenites, Cedripites,* and *Abiespollenites*; the Podocarpus pollen is common (genera *Podocarpidites, Dacrydiumites,* and *Phyllocladidites*); *Vitreispotrites pallidus* (Reisingen) Nilson is also present. The role of *Podozamites* is insignificant; representatives of genera *Araucariacidites, Ginkgocycadophytus, Ephedripites,* and *Gnetaceaepollenites* are occur sporadically.

Spores are diverse, and *Leiotriletes, Laevigatosporites,* and *Osmundacidites* dominate over *Sphagnumsporites, Lycopodiumsporites, Cyathidites,* and *Gleicheniidites*; selaginells are represented by rare grains of *Aquitriradites spinulosus* Cook. et Dett., *Selaginella utriculosa* Krasn., and *S. kemensis* Chl. et Krasn. Other taxa, such as *Rouseisporites laevigatus* Pocock, *Divisisporites euskirchenensis* Thompson, and *Cingulatisporites euskirchensoides* Delc. et Sprum., are less abundant; *Cicatricosisporites gorogensis* Pot. et Gell., *C. stoveri* Pocock., *C. tersus* (Bolch.) Chl., *Appendicisporites* sp., *Lygodiumsporites subsimplex* (Naum.) Bolch., *Concavissimisporites asper* (Bolch.) Pocock., *Taurocusporites reduncus* (Bolch.) Stover, *Foveosporites senomanicus* (Chl.) Swetz., and others are sporadic components.

The pollen of angiosperms is represented by tricolpate, rare tricolporate, and other forms like *Quercus sparsus* (Mart.) Samoil., Menispermaceae, Platanaceae, and *Quercus*.

Gymnosperms typical of the Middle Cretaceous dominate in the palynological assemblage studied. Spores are represented by individual specimens of the Albian–Cenomanian palynoflora, whereas pollen of typical Senonian angiosperms characteristic of the Enmyvaam palynological assemblage is absent in the Emuneret flora. Judging from the general taxonomic composition and the predominant pollen of the *Taxodiaceaepollenites–Cupressaceaepollenites* group, the Emuneret palynological assemblage is similar to the palynological assemblage identified in the lower part of the Kanguk Formation, on Ellesmere Island of the Canadian Arctic Archipelago (Nunes-Betelu and Hills, 1992), and referred to the late Turonian.

Nunez-Betelu, who reviewed the results of our analysis of palynological samples (personal communication, 1994), believes that the Enmyvaam palynological assemblage corresponds to the third sedimentation cycle of the Kanguk Formation, whose age is reckoned to be Santonian, because the upper Coniacian marine beds comprising *Scaphites depressus* Reeside (Hills *et al.,* 1994) are found at the base of the third sedimen-

tation cycle and the fourth sedimentation cycle referred to the Campanian begins with marine beds comprising radiate–costate inocerams.

The Merevaam palynological assemblage is identified from two samples taken from sedimentary rocks, which occur among basaltic flows of the upper part of the Enmyvaam Formation (Figs. 1, 2, Site 115). The samples are rich in microfossils (942 grains in the first sample and 1228 in the second); the pollen of gymnosperms dominates here (84.0–90.0%), spores make up 8.6–13.7%, and angiosperms represent 1.4–2.3%.

Despite a similar taxonomic composition of gymnosperms, the pollen ratio of different plant groups varies greatly. Sample 115-1 comprises forms among which diverse bisaccate forms dominate (64%) and the content of the pollen similar to that of Taxodiaceae (*Taxodiceaepollenites–Cupressaceaepollenites* sp., *Taxodiaceaepollenites* sp., *Taxodiumpollenites* sp.) averages 20%, whereas *Glyptostrobuspollenites* sp. and *Sequoiapollenites* sp. are rare. In Sample 115-3, taken 1.5 m upsection, the dominant Taxodiaceae (81.7%) associate with few bisaccate forms (8.3%).

Among less abundant spores, there are usually specimens of pisolitic fern *Laevigatosporites ovatus* Wilson et Webster; less abundant are spores of spiny *Osmunda granulata* (Mal.) Chl., *Osmundacidites wellmanii* Couper, *O. nicanicus* (Verb.) E. Sem., *O.* sp., *Baculatisporites* sp., and smooth triradiate *Leiotriletes* sp.; other specimens represent mosses *Sphagnumsporites antiquasporites* (Wilson et Webster) Patonie, *Sph.* sp., *Stereisporites* sp., and diverse selaginells, such as *Lycopodiumsporites clavatoides* Couper, *L.* sp., and *Retitriletes austroclavatidites* (Cook.) Dett.

Scanty pollen of angiosperms is represented mainly by tricolpate forms *Tricolpites* spp.; tricolporate forms with a smooth grain surface and retuculate, granulated, and tubercular ornamentation, such as *Tricolporopollenites* sp. and *Retitricolpites georgensis* Brenner, are rare. Pollen of *Triporopollenites* sp., *Liliacidites* aff. *creticus* N. Mtch., and forms similar to *Alnus* and Menispermaceae of the *Myrica* and *Platanus* types are rare.

Being less diverse, these spectra are similar in many respects to the Enmyvaam and Emuneret palynological assemblages. The spectra are free from forms characteristic of the latest Early–earliest Late Cretaceous, and from pollen of the supergroup *Triprojectacites* N. Match. developed in the early Senonian. It seems these are spectra of short-lived plants that appeared during the formation of a volcanic plateau. The taxonomic composition does not forbid and the stratigraphic position allows referring the Merevaam Formation to the early Senonian age.

MAGNETOSTRATIGRAPHIC DATA

During the field work of 1991 and 1993, a regular sampling of rocks of the OCVB was conducted for paleomagnetic analysis. The material collected has yet to be analyzed though the principal result of the study is quite evident and demonstrates that, among predominant rocks with normal magnetization, there are section segments and structures with reversed polarity (Raikevich, 1995).

A sequence of rocks (140–150 m thick) with reversed magnetization makes up the central part of the Emuneret Formation (Figs. 2, 3, units *em-5, em-6, em-7,* and *em-8*). This was observed in outcrops, and field observations were confirmed by laboratory studies. The sequence is similar in petrographic composition to extrusive and pyroclastic rocks of the Snezhnaya Caldera, though the structure of the caldera complex and stratified series of the middle part of the Emuneret Formation is different. The age correlation of these rocks by general geologic methods is impossible. Paleomagnetic data suggest that the Snezhnaya Caldera could be a source for the stratified rock series of the middle part of the Emuneret Formation.

The upper part of the formation begins with ignimbrites and vitroignimbrites (Unit *em-9*), which have the normal magnetic polarity. They are overlain by a member of tuffstones, tuff-siltstones, and tuffs of intermediate and basic composition (Unit *em-10*), in the lower part of which there is a packet 10 m thick displaying the reversed polarity.

The uppermost parts of the Emuneret Formation (the upper half of Unit *em-10* and Unit *em-11*) and all intervals of the Emuneret Formation sampled in the Emuneret River basin have the normal magnetic polarity.

In the Mechekrynnetveem River basin (Fig. 1, Site 8), a member of mudstones and siltstones is characterized by the reversed magnetic polarity. The member is overlapped by basalts with the reversed magnetization and a visible thickness of 15–20 m. Other basalts sampled in the Mechekrynnetveem structure have the normal magnetic polarity; their stratigraphic position with respect to siltstones, mudstones, and basalts with the reversed magnetization is not understood.

Thus, three zones of rocks with the reversed magnetization have been revealed at present in the stratigraphic sequence of volcanic deposits of the late evolutionary stage of the Central Chukchi sector of the OCVB. Volcanological studies show that the eruptive caldera formation was instant from the viewpoint of geologic time, and even a more extended process of caldera complex deposition could hardly have exceeded 30–40 k.y. (Braitseva *et al.*, 1995; Grib and Leonov, 1993). The eruption and cooling of basalt flows or a series of basalt flows totalling 20–30 m are also momentary events in the geologic sense. Thus, the reversed polarity zones revealed, as well as the normal polarity zone *em-9* located between them, mark episodes of extremely quick changes in the magnetic field of the Earth.

AGE OF VOLCANIC DEPOSITS OF THE LATE STAGE IN EVOLUTION OF THE CENTRAL CHUKCHI SECTOR OF THE OCVB

When discussing the age of volcanic deposits of the late stage in evolution of the OCSI, we prefer recent palynological data as more representative and unconflicting evidence. For instance, it is precisely the palynological studies that have shown that the upper boundary of the Arkagala flora should be referred to the early Senonian (Markevich, 1989; Belyi, 1994). Nevertheless, the lower boundary of this flora cannot be dated back to the Turonian or the beginning of the Senonian, as there is no evidence discrediting the position of the Gyrbykan (Arkagala) floral assemblage in the phytostratigraphic sequence of the northern part of the Ul'inskii trough, which was established by Lebedev (1987), and the lower boundary of the Arkagala flora should still be referred to the Cenomanian. This is also confirmed by a close phylogenetic association of the Arkagala stratoflora with the Chaun floral assemblage dated back to the late Albian.

The Ergyvaam Formation is assigned to the Cenomanian–Turonian on the basis of the representative Arkagala-type taphoflora enclosed in it and with regard to the stratigraphic position of the unit. Lebedev correlated the Ust'-Emuneret floral assemblage with the Barykov flora of Ugol'naya Bay and placed it at the level of the early Campanian (Lebedev, 1987). The basic, if not single, argument adduced to prove it is the presence of leaves of *"Quercus" tchukotica* Abram. in both taphofloras. Distinctions between the general taxonomic compositions of the taphofloras have not been taken into account, though they are essential. For instance, the Barykov flora (Herman and Lebedev, 1991) contains no mesophyte relicts, such as *Sphenobaiera* spp. and *Phoenicopsis* ex gr. *angustifolia* Heer, which are abundant in the Ust'-Emuneret assemblage. The Barykov flora is depleted in ferns and distinct because of a great diversity of angiosperms: the flora comprises *Magnoliaephyllum magnificum* (Dawson) Bell, *Platanus* cf. *marginata* (Lesq.) Heer, *Paraprotophyllum ignatianum* (Krysht. et Baik.) Herman, *"Quercus" viburnifolia* Lesq., *Grewiopsis akhmetjevii* Vachr. et Herman, *Araliaephyllum polevoii* (Krysht.) Krassil., *Rhamnites septentrionalis* (Krysht.), *"Vitis" stantoni* (Knowlton) Brown, *"Vitis"* ex gr. *populoides* Holl., and *Cissites beljaevii* Herman, which are all unknown in the Ust'-Emuneret floral assemblage.

As is evident from the foregoing, the general taxonomic composition of the Ust'-Emuneret floral assemblage indicates that it corresponds to an earlier stage of evolution of plants, in comparison with the Barykov flora. Palynological data suggest the late Turonian age of deposits, which comprise plant imprints of the Ust'-Emuneret floral assemblage. The rocks of the overlying middle part of the Emuneret Formation have the reversed polarity; this is the first R-zone established in volcanogenic deposits of the CCS and the OCVB as a whole. It is most likely that this zone is correlative with the R-zone revealed in the lower part of the Barykov Formation, whose layers correspond to the middle Coniacian *Inoceramus involutus* Zone (Pergament *et al.*, 1971). The reversed polarity zone is also established in the Coniacian deposits of Central Asia and Transcaucaia (Pecherskii *et al.*, 1983); in the general magnetostratigraphic scale of the Phanerozoic, this interval is referred to as the Klyuev R-zone (*Paleomagnitologiya,* 1982).

Thus, the evidence cited above offers a possibility to define the position of the boundary between the Turonian and Coniacian within the interval of stratigraphic units *em-3* and *em-4* of the Emuneret Formation section at the Enmyvaam River.

The Enmyvaam palynological assemblage occurs in various rock associations having a significant thickness in different structures of the Central Chukchi sector of the OCVB. The assemblage indicates an important terminal stage in the evolution of the Cretaceous flora in continental areas of northeastern Asia, when cenophytic conifers dominated over diverse, but less abundant, angiosperms. The leaf flora of that time (the Arkagala stratoflora) still comprised abundant *Spenobaiera, Phoenicopsis,* and *Czekanowsia,* but *Tchaunia, Heilungia,* and *Podozamites* became extinct. The beginning of this stage is referred probably to the Cenomanian.

The Enmyvaam palynological assemblage at the Enmyvaam River is established in the lower beds of the Enmyvaam Formation, the rocks of which are characterized by the normal magnetization. Taking into account that the R-zone of the Emuneret Formation corresponds to the Klyuev R-zone (the lower Coniacian), the lower rock series of the Enmyvaam Formation can be referred to the end of the Coniacian–the beginning of the Santonian.

The Merevaam palynological assemblage of low stratigraphic significance cannot be younger than the early Senonian. It was formed simultaneously with the lowermost segment of the upper part of the Enmyvaam Formation, whose lavas are characterized by the normal magnetic polarity throughout the 200-m thick interval.

According to the scale of the magnetic field inversion, the Santonian crowns a long period of normal polarity in the Middle Cretaceous, and the first major R-zone as long as 4 m.y. falls within the early Campanian (Harland *et al.*, 1982). Thus, it seems reasonable to confine the Enmyvaan Formation to the Santonian, which spans the optimal time interval of 4.5 m.y. for the 1000-m sequence of plateau basalts to accumulate. However, according to recent studies of the Cretaceous deposits in Japan, the first major R-zone falls within the middle Campanian, and its lower part, similar to the Santonian beds, is characterized by normal polarity (Toshimitsu *et al.*, 1995). Therefore, taking into

account the data considered above, the uppermost parts of the Enmyvaam Formation can be referred to the Campanian as well.

In their external features (thin, persistent, extended flows and a small amount of amygdaloidal rocks, but widespread vesicular lavas), basalts of the Mechekrynnet volcanostructure resemble those from the upper part of the Enmyvaam Formation observed at Merevaam River. It should be added that soaking siltstones and mudstones at Mechekrynnetveem River greatly resemble thin sticky clays and soaking coaly siltstones from the lower part of the Enmyvaam Formation (Basalt Cliffs). As was discussed above, the reverse polarity zone of basalts exposed in the Mechekrynnetveem River basin (their visible thickness is 20 m) occur above siltstones and mudstones. The basalts likely recorded a very short reverse polarity episode, which has not been established in sections of the Enmyvaam River because they may be incomplete, or this interval could have been missed in the course of sampling. Fig. 2 demonstrates this reverse polarity zone at the base of the upper part of the Enmyvaam Formation indicated by a question mark (?).

Earlier, Raikevich (1995) suggested a quite different age interpretation of the magnetostratigraphic material presented in this work. He correlated the R-zone of the middle part of the Emuneret Formation to the early Campanian R-zone of the general magnetostratigraphic scale (Harland *et al.*, 1985) and referred the upper part of the Emuneret Formation, together with the Enmyvaam Formation, to the upper Campanian and Maastrichtian. Such an interpretation contradicts the palynological data presented in the work. Besides, in Belyi's opinion, the interpretation is inconsistent with data on the geology of the region studied and with the current understanding of the evolution of volcanic structures.

Tuffs, ignimbrites, and vitrophyres with reversed magnetization from the middle part of the Emuneret Formation represent part of a unified succession of volcanic deposits composing this unit as a whole. No hiatus is observed between ignimbrites and tuffs with reversed magnetization and similar rocks with normal magnetization underlying and overlying them. As was discussed above, volcanic deposits in the R-zone of the middle part of the Emuneret Formation were formed as a result of one or several caldera-forming eruptions contiguous in time, i.e., during a very short time interval. Therefore, taking into account the compositional uniformity and continuity of the Emuneret Formation, the R-zone considered cannot be referred to the Campanian reversed polarity zone, the most prolonged (about 4 m.y.) in the Late Cretaceous.

Formation of a basaltic plateau is an overwhelmingly more extended and quiescent process than an ignimbrite eruption. Modern study of Deccan traps showed that eruption of lavas of this immense volcanic province lasted 3 m.y. (Deshmukh, 1990; Venkatesan and Pande, 1990). Thus, the Santonian is a more than sufficient interval (4–5 m.y.) for accumulation of the 1000-m-thick sequence of normally magnetized plateau basalts of the Enmyvaam Formation.

It is of interest to compare the inferred ages of stratigraphic units of the late evolutionary stage of the Central Chukchi sector in the OCVB with the duration of the Cretaceous stages shown in "A Geological Time Scale" (Harland *et al.*, 1985). The scale allocates 6.5 m.y. for the Cenomanian, which is subdivided into three substages (arbitrarily of equal duration); 2.5 m.y. for the Turonian, also subdivided into three substages; 1 m.y. for the Coniacian (two substages), 4.5 m.y. for the Santonian (two substages), and 10 m.y. for the Campanian (two substages). The interval from the beginning of the Cenomanian (97.5 m.y. ago) to the end of the Santonian (83 m.y. ago) is 14.5 m.y. Estimated accumulation periods of the Emuneret (late Turonian–Coniacian) and Enmyvaam (Santonian–the beginning of the Campanian) formations are 1.8 and 4.5–5 m.y., respectively. The Ergyvaam Formation is unlikely to have formed at the beginning of the Cenomanian. Assuming that this happened during the late Cenomanian and early and middle Turonian, we obtain the approximately summary period of 4 m.y. The evaluated periods, when the Ergyvaam, Emuneret, and Enmyvaam formations were formed, are in good consistency with their relative volumes, and this lends credence to our estimates. Thus the total length of the late evolutionary stage of the CCS in the OCVB amounts to 10 m.y.

CONCLUSION

Volcanic deposits of the late stage in CCS evolution represent natural formations structurally isolated in the OCVB, whose stratigraphy and age were studied using the paleobotanical, palynological and paleomagnetic methods.

The combination of these methods offers a good opportunity to study the successive accumulation of volcanic deposits and their stratigraphic correlation. The palynological method, which was formerly not widely used in the study of the stratigraphy of the Cretaceous volcanogenic deposits, is of special significance. At present, palynological studies have completely stopped in northeastern Russia. This is only one of a series of consequences resulting from commercial capitalistic practice in Russian geology.

Reviewer A.B. Herman

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Elementary Stratigraphic Units

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Abstract—It has been established that every tenth bed of the Upper Cretaceous–Danian flysch deposits can be traced along the southern slope of the Greater Caucasus over a distance of 1035 km. This allowed 196 synchronous markers to be established in this region. Their synchronism is based on a relative number of cyclites and confirmed by paleontologic data. The formation time of one global stratigraphic unit is assumed to average 22 ± 9 k.y.

Key words: *stratigraphic unit, stratigraphy, stage, age, phase, member, package, bed, laminae, Greater Caucasus, Upper Cretaceous, flysch, carbonate formation.*

GENERAL INFORMATION

Two centuries ago, the prevalent stratigraphic method that dealt with lithostratigraphic units, including the global Primary, Secondary, Tertiary, and Diluvium rocks (Arduino, 1750), gave way to the new biostratigraphic method, which quickly conquered the minds of geologists. At present, this method is a basic one in studies on Phanerozoic deposits. However, its potential is not unlimited. It is useless when applied to fossil-free deposits. The smallest biostratigraphic units, such as chains, gilberts (Gilbert, 1895), and events, comprise not only a number of beds but sometimes subformations and even formations.[1] For instance, upper Campanian deposits in the Novorossiisk reference section of the northwestern Caucasus comprise 4 formations, 10 subformations, 74 stratigraphic units of the Greater Caucasus scale (Afanas'ev, 1993), 4059 natural cyclites, and more than 10000 beds.

From the aforesaid, it seems reasonable to return to the lithostratigraphic method when subdividing ancient rock formations. To define the potential of this method, a thorough study on Upper Cretaceous–Danian sequences of the Greater Caucasus was conducted over a long period of time (1953–1994). The investigation was performed in two stages. First, sedimentary rock varieties were classified according to their origin (clastic, biogenic, volcanogenic), granulometry, carbonate content, silica content, coloration, and other properties (Afanas'ev, 1993). Second, following Vassoevich (1948), these varieties were grouped into natural cyclites (C) and their elements (CE): (1) *clastic* avalanche and landslide breccias of the one-prime cyclite element (0^1 CE); olistostromes (0 CE); relatively slow-forming conglomerates, sandstones, and siltstones often displaying horizontal microlamellar structure (I^1 CE); fast-forming graded sandstones and siltstones (I CE); and siltoliths, such as marls, claystones, and limestones (II CE); (2) *biogenic* limestones, marls, and products of their dissolution, for instance, lutites and clays (III CE); (3) *volcanogenic* tuffs, tuffites, bentonite clays, and clays (IV CE). Third, more than one million beds in more than 300 sections, first of all those located in the main Novorossiisk, Chiauri, Kobystan, and Dagestan depressions of the Greater Caucasus, were measured bed-by-bed with an accuracy up to one millimeter. All particular sections represent stratigraphic units with the strict or Stenson's bed succession. The analysis of composition, thickness, and relationships of different beds allowed geologists to define groups, formations, subformations, members, packages, and marker beds in the above-mentioned depressions of the Greater Caucasus and to assign these units to stages and substages. Recent stratigraphic schemes of the Upper Cretaceous–Danian of the Greater Caucasus can be found in the following summarizing publications and separate papers: (*Geologiya* ..., 1968; *Stratigrafiya* ..., 1986; *Verkhnii* ..., 1986; Afanas'ev, 1987b, 1993; Gambashidze, 1973; Khalilov and Ali-Zade, 1968; and others). Stratigraphic schemes of the Novorossiisk, Chiauri, Kobystan, and Dagestan depressions were correlated at the formation, subformation, and member levels based on paleontologic data (Afanas'ev, 1993).

The elementary Upper Cretaceous to Danian stratigraphic units common to all the depressions of the Greater Caucasus could be defined only on the basis of bed-by-bed correlation, and the problem to be solved first was the tracing of beds over a distance of hundreds and the first thousands of kilometers.

BED-BY-BED CORRELATION

The problem of bed-by-bed correlation has attracted researchers for almost a century (De-Geer, 1910). Vassoevich (1948, p. 146) wrote that

> at the 11th Session of the International Geological Congress held in Stockholm, 1910, De-Geer

[1] The polarity event corresponds to a period not exceeding 100 000 years (*Glossary* ..., 1978).

put forward the idea on the possible correlation (connection) of diagrams compiled for the varved clays of Sweden and North America. In 1924, the Geochronological Institute was organized in Stockholm to coordinate all the research on teleconnection (i.e., connection over far distances). Shortly before that event (1920), De-Geer visited North America with the aim to correlate *varves* of the New and Old Worlds. In his article published in the *Sweden Geological Journal* in 1921, he correlated varves over a distance of 6250 km."

Vassoevich (1948) conducted a bed-by-bed correlation of upper Maastrichtian flysch sections of Kakhetia over a distance of 24 km.

Geologists face the correlation problem when studying an outcrop and observing dislocations that disturb the continuous strata succession. The comparison of strata lithology and thickness in bed members located in opposite sides of the fracture plane commonly allows their bed-by-bed correlation to be easily performed. The correlation reliability increases with an increase in bed number in these members. For a reliable correlation, their number should not be less than five; a desirable number of beds is several dozens (Afanas'ev, 1993).

When conducting bed-by-bed correlation using the method of De-Geer and Vassoevich, a geologist faces the problem of the compatibility of beds first of all in their thickness. To determine the compatibility degree, flysch members many meters thick were traced along their strike in several sections. In doing so, every bed was visually traced from one section to another over a distance of four kilometers (Vassoevich, 1948; Afanas'ev, 1961, 1967, 1993). This procedure revealed the principal possibility of bed-by-bed correlation over the first kilometers. When the distance between sections was 5–20 km, aerial photographs were used for their bed-by-bed correlation (Afanas'ev, 1967). When the distance was 30–200 km, sections several kilometers thick (Anapa, western Ozereika, Bzhida, and others) were studied with due regard for marker boundaries of formations and subformations. When the distance between sections exceeded 200 km, the whole Upper Cretaceous interval was investigated, and stage and substage boundaries defined by paleontologic data were taken into account.

Not all the beds can be traced over a distance of many tens and hundreds of kilometers even when studying flysch deposits. Some of them show facial variability; others wedge out or are disturbed by intraformation erosional surfaces. More stable are ash layers, large cyclites, and sandstone and marl beds, i.e., strata whose formation was related to volcanic activity, major earthquakes, and tsunami. These beds are traced in adjacent structural–formational zones, sometimes across the depression strike. For instance, some beds can be observed in the southern slope of the Greater Caucasus, on the one hand, and in Dagestan, the North Caucasus, on the other.

The bed continuity in a flysch formation is much more stable than that in other terrigenous formations. For instance, in the Riphean terrigenous rocks of the Strel'naya Gora Formation of the Turukhansk uplift (Ivanovskaya and Afanas'ev, 1970) where the 0.5-m-thick member was traced over 10 m, layers often wedge out and facially change even over a distance of 0.5 m, thus demonstrating their extreme discontinuity.

Let us consider the examples of continuous layers in Upper Cretaceous–Danian sections in the southern slope of the Greater Caucasus.

Earlier, the correlation of the 30-m-thick member of the middle Sukko Formation was conducted over a distance of 177 km in five sections of the Novorossiisk depression (Afanas'ev, 1961). The member comprises 96, 85, 59, 68, and 156 beds, including the Karadzhyuzlyu and Chukhuryurd markers (Afanas'ev, 1993), i.e., thick sandstone beds in the Anapa, western Ozereika, Myskhak, Bzhid, and Shepsa flysch sections, respectively.

All the sections show the presence of 35 regional ("West Caucasian") and tens of local beds that wedge out in various sections. If we denote the ratio of the bed number in the most complete section to that traced in two correlated sections by the letter "K," then we can observe the positive correlation between logarithms of K and L (the distance between correlated sections in kilometers).

In the 1960s, the correlation of the 12-m-thick Limanchik Member that occurs in the basal part of the upper Danian Tsitse Formation was conducted over a distance of 209 km in seven sections of the Novorossiisk depression (Afanas'ev, 1967). This member comprises 77, 62, 51, 46, and 48 beds in the western Ozereika, Anapa, Praskoveevka, Kadosh, and Golovinka sections, respectively. All the sections also incorporate 14 regional (in terms of the western Caucasus scale) and tens of local beds that wedge out in various sections. The positive correlation between log$K$ and log$L$ is also characteristic of these sections (Bol'shev and Smirnov, 1965).

Finally, in the 1980s, the correlation of the 80-m-thick Abrau Member of the lower Snegurevka Subhorizon (upper Maastrichtian) of the Greater Caucasus was conducted over a distance of 1035 km (Afanas'ev, 1987b). The member comprises 410 beds in the western Ozereika section of the Novorossiisk depression and 325 beds in the Agdarachai section of the Kobystan depression. Both sections have 27 beds in common. In this case, the log$K$ is equal to log410/27 = 2.182, and log$L$ is equal to log1035 = 3.015.

All three above-mentioned groups of coupled log$K$ and log$L$ values are close to each other. According to these data, the average

$$\log K = 0.51 \log L - 0.53. \quad (1)$$

In accordance with equation (1), all beds of flysch deposits can be traced in the southern slope of the Greater Caucasus over a distance of 8–14 km; every

second bed, over 50 km; every third bed, over 100 km; every fourth bed, over 200 km; every fifth bed, over 500 km; and every tenth bed, over 1035 km. As the Novorossiisk stratotype section of the upper Cretaceous–Danian deposits comprises 23046 cyclites, or 60 840 beds, whereas the Dagestan section incorporates only 2044 cyclites, or 4300 beds, every bed of the latter corresponds to 14 beds in the former. When applying equation (1) for correlation of the Novorossiisk and Dagestan sections separated by a distance of 1035 km, one can calculate that 5850 of the total 60 840 beds appear to be common (in terms of the Greater Caucasus scale) for both sections; i.e., every tenth bed of the Novorossiisk section and *every* bed of the Dagestan section are correlative. Accepting the duration of the Late Cretaceous–Danian as $T_1 = 96253 - 62216 = 34037$ k.y., we can calculate the average duration of the time interval corresponding to the formation of a single stratigraphic unit of the Greater Caucasus: $T_2 = 34037 \times 10.4/60840 = 5.818$ k.y., which corresponds to an "episode," or the duration of the eleventh-class cyclite formation (Afanas'ev, 1993, Table 26), i.e., to the formation time of a standard submember. Beginning in 1957, the submember correlation diagrams that were compiled on the basis of eleventh-class cyclites were widely used in field and laboratory studies of the Novorossiisk sections of the northwestern Caucasus (Afanas'ev, 1961; Afanas'ev *et al.*, 1961).

Thus, when correlating Upper Cretaceous–Danian deposits in the four main depressions of the Greater Caucasus (Novorossiisk, Chiauri, Kobystan, and Dagestan depressions), 5850 stratigraphic units of the Greater Caucasus scale corresponding to every bed of the Dagestan reference section can be theoretically defined.

ELEMENTARY STRATIGRAPHIC UNITS

Prior to the formulation of an empirical equation and related conclusions, the Upper Cretaceous–Danian deposits in the four main depressions of the Greater Caucasus (Novorossiisk, Chiauri, Kobystan, and Dagestan depressions) were correlated by using the analysis of markers (Afanas'ev, 1987b, 1993).

Solution of lithostratigraphic problems is based on *unique lithology and thickness* of stratigraphic units that were formed at *certain moments* of geological history. The clay bed with an elevated iridium content in basal Danian deposits can serve as an example of such a unit.

The bed-by-bed examination of sections and identification of all their beds more than 0.5 mm thick significantly increases the information value and, thus, decreases the possibility of erroneous correlation between them. In my previous (Afanas'ev, 1987b, 1993) and present works, detalization of stratigraphic schemes of main depressions of the Greater Caucasus was performed based on the identification of marker members, packages, cyclites, and beds.

Marker lithostratigraphic units comprise all unusual strata that differ from adjacent "usual" deposits, first, in their *greater thickness* and, less commonly, in unique lithology (e.g., bentonite clays, ankerite beds, etc.). To evaluate objectively the marker significance, the following rule was accepted: (1) when the marker bed is unique in the Upper Cretaceous–Danian section and characterized by the greatest thickness, it is considered to have "complete" rank with the index $I_{01} = 480$.[2] When the considered marker is second among the thickest Upper Cretaceous–Danian units, it is denoted by the index $I_{02} = 480/2 = 240$, and, when it is the third, by the index $I_{03} = 480/3 = 180$, etc. As the Upper Cretaceous–Danian deposits in the Novorossiisk depression comprise four rock groups, the largest marker of any of them has the index $I_{11} = 480/4 = 120$. Every group commonly consists of four formations; therefore, the largest marker in any of them is denoted by the index $I_{21} = 120/4 = 30$. Every formation, in turn, commonly comprises three subformations, and, thus, the largest marker in any subformation has the index $I_{31} = 30/3 = 10$. If the subformation marker is tenth among the thickest units, it is denoted by the index $I_{310} = 10/10 = 1$. When the marker member (submember, cyclite) includes several marker beds with pertinent indices, these latter are summed.

The synchronous lithologic marker unit is formed during the same *short interval* of geological time. We should neglect here the small temporal differences (day, week) related to the velocity of the distribution of sedimentary material that composes the marker unit (ash, turbidite material, and others). The synchronism of formation is determined based on the thickness scale transformed into the *time scale.* After Artur Holmes' works (Holmes, 1913), lithologic criteria, for instance, the number of natural cyclites in the Novorossiisk reference section (Afanas'ev, 1987a), are widely used for interpolation of isotopic age determinations. Let us take cyclites of the Novorossiisk reference section as *relative time units.* The relative geological age of any marker unit in the considered section is determined by the ratio (T%) of the natural cyclite number (NC) in the interval from the base of the marker to the base of the considered lithostratigraphic unit to the total NC number in the latter. When a particular marker unit is synchronous in different sections, the relative NC number (T_i%) is expected to be the same in all the sections. In fact, the T_i% values in them coincide only at a first approximation, differing from each other (more exactly, from the average T_i% value) by the value of $\Delta_i T$% used to calculate the mean square deviation σ_0% from the NC number in the considered lithostratigraphic unit. The average value of σ_0 is 3.13%.

[2] $480 = 4 \times 4 \times 3 \times 10$, where 4 is the number of suprahorizons in the Caucasus Superhorizon, 4 is the number of horizons in the suprahorizon, 3 is the number of subhorizons in the horizon, and 10 is the number of thick beds in the subhorizon.

Let us consider a particular example. The Caucasus Superhorizon in the Greater Caucasus includes abundant limestone beds whose share is twenty times higher than in underlying and overlying superhorizons; its carbonate content is five times higher, and its clay content five times lower (Afanas'ev, 1993). In addition, the base of the Caucasus Superhorizon is also marked by the manifold increase of the calcareous marl and clayey marl share. In flysch deposits of the southern slope of the Greater Caucasus, the share of these rocks decreases by two orders.

Based on paleontologic evidence, the age of the Caucasus Superhorizon was determined: its basal layers were dated as the earliest Cenomanian, and the top, as the terminal Danian (Afanas'ev, 1993).

The best lithologic marker in the Caucasus Superhorizon is represented by the Divnomorsk Member (2–15 m; $I_{01} = 480$), which incorporates two bentonite clay beds, one of which is the thickest and ranked by $I_{01} = 480$ in all the considered depressions; the other is in second–sixth place among the bentonite clay beds in the Caucasus Superhorizon, being ranked by values of 240, 160, and 80 in the Kobystan, Dagestan, and Novorossiisk depressions, respectively (in the Chiauri depression, the bentonite clay bed is not exposed).

The Caucasus Superhorizon in the Novorossiisk depression begins with the first NC and terminates with the 23088th NC; the base of the Divnomorsk Member with bentonite clays corresponds to the base of the 12889th cyclite (figure). In other depressions of the Greater Caucasus, the cyclite number in the Caucasus Superhorizon is different: 11519, 16372, and 2044 in the Chiauri, Kobystan, and Dagestan depressions, respectively. Therefore, in the Dagestan depression, 12889 cyclites are missing, and the base of the Divnomorsk Member corresponds there with the base of the 819th cyclite. This means that sections should be correlated based on the relative cyclite number rather than on their total number. The lower boundary of the Divnomorsk Member is located (in terms of geological time) at the level of $T_{nov} = 55.8\%$ from the base of the Caucasus Superhorizon according to the number of cyclites in the Novorossiisk depression and at $T_{chi} = 45.1\%$ in the Chiauri, $T_{kob} = 58.4\%$ in the Kobystan, and $T_{dag} = 59.9\%$ in the Dagestan sections. T_{chi} sharply differs from all other T values (58.0% in average), which is related to major hiatuses in the pre-Divnomorsk part of the Caucasus Superhorizon in the Chiauri section. Assuming that 42.0% of the post-Divnomorsk part of the Caucasus Superhorizon (the average value for the Novorossiisk, Kobystan, and Dagestan sections) corresponds to 54.9% of the same part of this unit in the Chiauri section, it is easy to estimate the share of missing deposits (W) in the latter: $W = (1 - 42.0/54.9) \times 100\% = 23.5\%$. Deviations from $T = 58\%$ are as follows: –2.2% in the Novorossiisk, 0.4% in the Kobystan, and 1.9% in the Dagestan sections. The mean square value of ΔT is 1.69%. Is this value high or low? Does this ΔT value evidence the synchronous formation of the Divnomorsk Member in the Greater Caucasus sections or not?

Let us consider Table 5.2 published by Bol'shev and Smirnov (1965). This table may be used for calculation of the mean square deviation ΔT at the confidence level of 5% ($Q = 5\%$, $P = 0.95$). According to this table, the Divnomorsk Member in the Novorossiisk depression is characterized by $\Delta T_{nov} = 0.0027$, $\Delta T_{kob} = 0.0032$, and $\Delta T_{dag} = 0.0090$. All three ΔT_i values are, respectively, six, five, and two times lower than $\Delta T = 1.69\% = 0.0169$. This fact indicates their incompatibility and suggests additional causes of the ΔT difference. First and foremost, these are nondeposition and intraformational hiatuses.

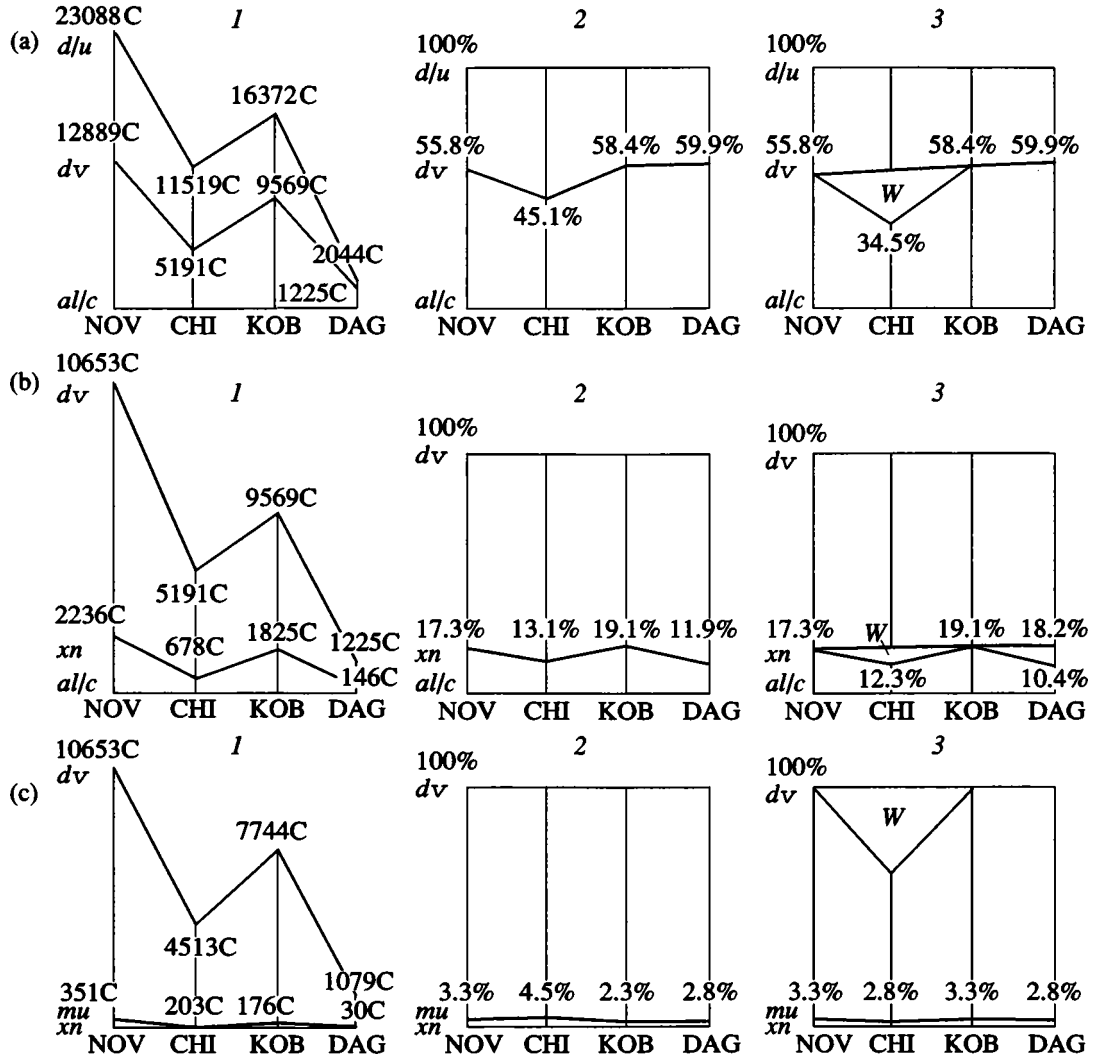
In one of my previous articles (Afanas'ev, 1993), I noted 49489 hiatuses confined to the base of all cyclites of the Caucasus Superhorizon in the Novorossiisk, Chiauri, Kobystan, and Dagestan depressions of the Greater Caucasus. Thirty-two or 0.1% of them are apparent. They were defined based on the absence of 182 to 2224 cyclites in these sections and are commonly confined to the base of formations corresponding to cyclites of the seventh class spanning time intervals of 4 m.y., and also to subformations that represent cyclites of the eighth class with time intervals of 1 m.y. Six hundred ninety-seven hiatuses (or 1.4% of their total number) are mainly cryptic and were defined based on the sharp ΔT deviation due to the absence of 12–180 cyclites. These cryptic hiatuses are confined to the base of members, which are cyclites of the ninth class corresponding to the time intervals of 200 k.y. (Afanas'ev, 1993). Another 5513 hiatuses (or 11.1% of their total number) are mainly assumed and were defined based on significant ΔT deviations due to the absence of 1–11 cyclites. They are commonly confined to the base of submembers, the cyclites of the eleventh class corresponding to time intervals of 40 k.y. (Milankovich cycles). Most hiatuses, namely 87.3% of their total number, affect only the upper part of cyclites primarily composed of biogenic deposits.

The second cause of the ΔT and $\Delta^1 T$ discordance is related to the limited extension of beds and stratal concretions, which wedge out along the strike. Facies changes are the third cause of respective disagreements.

Thus, the Divnomorsk Member can be considered, with certain reservations concerning probable hiatuses and intraformational erosion, as synchronous in all depressions of the Greater Caucasus and representing the marker unit for the entire region.

According to paleontologic data, the Divnomorsk Member in all four depressions of the Greater Caucasus corresponds to the uppermost Campanian Stage (Afanas'ev, 1993; *Verkhnii* ..., 1986; and others).

The Caucasus Superhorizon is divided by the Divnomorsk Member into the pre-Divnomorsk unit, which comprises 12889 cyclites in the Novorossiisk reference section, and the post-Divnomorsk unit incor-



Correlation of sections of the Greater Caucasus. (a) Cenomanian–Danian Caucasus Superhorizon: (*d/u*) the Danian–Unnamedian boundary (Afanas'ev, 1993, 1994); (*d/v*) Divnomorsk Member with bentonite clays; (*al/c*) the Albian–Cenomanian boundary; (C) cyclites, (*W*) hiatuses and intraformational unconformities; (Nov) Novorossiisk reference section; (CHI) Chiauri reference section (Georgia); (KOB) Kobystan reference section (Azerbaijan); (DAG) Dagestan reference section. (b) Cenomanian–Turonian Lipki–Divnomorsk stratigraphic unit: (*xn*) the Ksani Member with cherts and bentonite clays. (c) lower Turonian–upper Campanian Ksani–Divnomorsk stratigraphic unit: (*mu*) the Mugi package with bentonite clays. *1*, by the natural cyclite number (C); *2*, by the relative number of natural cyclites (*T*, %); *3*, by the relative number of natural cyclites including hiatuses (*W*).

porating 10199 cyclites of the same section. The "useful" or *pragmatic* marker is confined to the pre-Divnomorsk unit that is characterized by the higher number of cyclites. This marker is the Ksani or well-known middle Ananuri Subhorizon (Rengarten, 1924; Karstens, 1932; *Geologiya ...*, 1969; *Stratigrafiya ...*, 1986; Afanas'ev, 1993). The Ksani Subhorizon in the Novorossiisk, Chiauri, and Kobystan sections is represented by a unique siliceous member ranked by the index value of 480. In the Chiauri and Dagestan sections, it incorporates three and two thick bentonite clay beds with $I = 240 + 160 + 48$ and $I = 240 + 69$, respectively: in the Kobystan section, there is a single bentonite clay member with $I = 48$. When

$$I > 0.00326N + 0.0208n, \qquad (2)$$

the marker unit is classified as nontrivial. In this equation, N and n correspond to the cyclite number in the considered superhorizon and marker unit, respectively.

The considered example shows that $I_{nov} = 480$, $I_{chi} = 928$, $I_{kob} = 528$, and $I_{dag} = 309$. The value of $I_{chi} = 928$, which is probably dependent on some local factors, should be excluded. The average value I_{ave} for the remaining three indices is 428. As follows from (2), I_{ave} is > 45.7, which means that the Ksani marker unit is of the nontrivial type.

The basal part of the Ksani Member is characterized by the following parameters: $T_{nov} = 17.3\%$, $T_{chi} = 13.1\%$, $T_{kob} = 19.1\%$, and $T_{dag} = 11.9\%$ (figure). The two lowest values of T_{chi} and T_{dag} are probably indicative of hiatuses (*W*) in the pre-Ksani part of the consid-

ered stratigraphic unit: $W_{chi} = (1 - 81.8/86.9) \times 100\% = 5.9\%$, $W_{dag} = (1 - 81.2/88.1) \times 100\% = 7.8\%$, and the average value of $T_{nov + kob}$ is 18.2%.

The mean square deviation from the average value is $\sigma = 1.8 \times 0.8862 = 1.60\%$, which is almost two times less than $\sigma_0 = 3.1\%$, and may suggest a similar age for the base of the Ksani marker unit in the Novorossiisk and Kobystan reference sections. Accordingly, this unit represents by its rank the marker of the Greater Caucasus scale.

According to paleontologic data, basal layers of the Ksani Member in all four sections of the Greater Caucasus correspond to the earliest Turonian (Afanas'ev, 1993; *Verkhnii* ..., 1986).

Thus, the Caucasus Superhorizon is divided by two markers into three members: (1) the pre-Ksani unit (2226 cyclites in the Novorossiisk reference section); (2) the Ksani–Divnomorsk unit (10653 cyclites in the same section); and (3) the post-Divnomorsk unit (10199 cyclites in the same section). The third pragmatic marker is located in the Ksani–Divnomorsk Member with the maximum cyclite number. This marker is represented by the Mugi Submember with bentonite clays (figure). In the northwestern Caucasus, the markers correspond to the submember (with the rank index 96) and a separate bed of bentonite clay (rank index 50) in the Novorossiisk reference section (total rank index 156), to the submember (rank index 480) and a separate bed of bentonite clay (rank index 160) in the Chiauri reference section (total rank index 560), and to beds of bentonite clay (rank index 240 + 160) and chert (rank index 160) in the Dagestan reference section (total rank index 560). If we exclude from consideration the Chiauri reference section with the highest rank, which is probably related to local factors, then the average value for the remaining three sections will be $I_{ave} = (156 \times 132 \times 560)^{1/3} = 226$. From (2) it follows that I_{ave} is >45.7 and again indicates the nontrivial character of the Mugi marker. The Ksani–Divnomorsk unit of the Novorossiisk reference section incorporates 10653 cyclites, which is much more than the Ksani–Mugi unit with 351 cyclites; their ratio is $T_{nov} = 3.3\%$. In the Chiauri, Kobystan, and Dagestan sections, this parameter is $T_{chi} = 203/4513 = 4.5\%$, $T_{kob} = 176/7744 = 2.3\%$, and $T_{dag} = 30/1079 = 2.8\%$, respectively. Among them, the T_{chi} value is outstanding, probably because of hiatuses: $W_{chi} = (1 - 2.8/4.5) = 37.8\%$. The $T_{nov + kob + dag}$ is 2.8%, and the mean square deviation $\sigma = 0.59\%$, is significantly lower than $\sigma_0 = 3.13\%$, thus indicating the similar geological age of the Mugi unit in the Novorossiisk, Kobystan, and Dagestan sections. The major hiatus in the Chiauri depression is responsible for the disappearance of the lower and middle Akheyany subhorizons (Afanas'ev, 1993). Thus, with its rank and age similarity, the Mugi unit can be referred to the markers of the Greater Caucasus scale.

Paleontologic data evidences that the Mugi unit with bentonite clays corresponds to the lower Turonian Substage.

Thus, the correlation procedure includes the following steps: (1) the marker of the maximum rank (I) should be defined in the considered stratigraphic unit; (2) $T_i\%$ values must be calculated for all reference sections; (3) the average T value ($T_{av}\%$) and deviations from the latter ($\Delta_i T\%$) are the next parameters to be calculated; (4) the mean square deviation $\sigma_x\%$ must be evaluated; (5) when $\sigma_x < 2\sigma_0$, the correlation validity is to be estimated (in other words, synchronism of the relative age of the marker may be assumed valid at the confidence level $Q = 5\%$); (6) with equation (2), the significance of the marker for the entire Greater Caucasus should be validated; and (7) the relative geological age must be checked by faunal and floral fossils.

Following this procedure, another 193 useful markers of the Greater Caucasus rank were defined and analyzed, including the lower Turonian Lysaya Gora cherts, upper Danian Limanchik Member with sandstones, lower Campanian Dzhankhot Member with bentonite clays, upper Maastrichtian Palagina Shchel' marl, upper Turonian Tkhab Member with olistostromes, upper Turonian Bakanka Member with sandstones, lower Maastrichtian Rybatskaya limestone, and others.

The age of all these markers up to the substage level was confirmed by paleontologic data.

With the increase in the Greater Caucasus marker significance and decrease in the cyclite number in the considered section interval, the similarity between primary values calculated using equation (2) and data from Table 5.2 (Bol'shev and Smirnov, 1965) constantly grows, and this suggests the absence of important intraformational hiatuses. At the same time, the presence of hiatuses that were previously established based on missing marker units was confirmed. In other words, hiatuses that were assumed in accordance with equation (2) acquired the rank of distinct missing markers.

The number of the defined Greater Caucasus markers (195) may be theoretically increased to at least 2347, i.e., to the number of natural cyclites in the Dagestan reference section (Afanas'ev, 1993). A principally similar conclusion was drawn by Cotillon (1989), who established that rhythmic patterns of Cretaceous pelagic deposits are the results of global events.

Equation (1) can be used to forecast the number of global ($L = 20000$ km) stratigraphic units. In my previous paper (Afanas'ev, 1993), I considered flysch deposits of the Snegurevka and Tsitse formations, which incorporate 4300 cyclites or 11395 beds in the Novorossiisk reference section. The formation period of these deposits lasted $T = (67441 \pm 26t) - (62225 \pm 24t) = 5215 \pm 35t$ k.y.; this parameter for one cyclite is equal to $12312.8 \pm 8.1t$ years, and for a single bed, to $457 \pm 3.1t$ years (t here denotes the quantile of normal distribution). According to equation (1), we may expect

only one global stratigraphic unit consisting of 47.4 ± 20.6t beds that can be traced over a distance of 20000 km and can be present in Upper Cretaceous–Danian flysch deposits of the Novorossiisk reference section. Thus, the formation time of a single standard stratigraphic marker is (47.4 ± 20.6t) × (457.7 ± 3.1t) years = 21693 ± 9428t years, which corresponds to the formation time of one gilbert (Gilbert, 1895), or one Milankovich cycle, the smallest among the four distinguished (Milankovich, 1941; Berger *et al.*, 1989).

Thus, the standard scale can include 569700/21.693 = 26262 Phanerozoic global stratigraphic units, which is 108 times more than their maximum number mentioned in my previous paper (Afanas'ev, 1994).

The number of global Phanerozoic stratigraphic units constantly increases. For instance, Arduino (1759) defined four such units (Primary, Secondary, Tertiary, and Diluvium). By 1841, all geological systems and, partly, series and stages were established. In 1987, I predicted that elaboration of the Phanerozoic geochronologic scale with 146 isotopic age data was expected by 1990 and with 264 by 2000 (Afanas'ev, 1987). Harland *et al.* (1990) published 124 age data for boundaries between Phanerozoic stratigraphic units, whereas in my paper (Afanas'ev, 1994) I listed 244 age determinations confirming my forecast of 1987. In accordance with the equation

$$\log n = 0.3058(T - 1970) + 1.5541, \qquad (3)$$

we may expect that by 2059–2072 (T = 2059–2072), the number of published age data for Phanerozoic boundaries will be $n = 26262 \pm 14072t$, and, thus, elaboration of the global Phanerozoic geochronologic scale will be accomplished.

CONCLUSION

From the analysis of data considered here and before (Afanas'ev *et al.*, 1961; Afanas'ev, 1967, 1987b), it follows that the use of the above-discussed method allows the bed-by-bed correlation of Upper Cretaceous–Danian flysch deposits in the southern slope of the Greater Caucasus over a distance of 1035 km. It was established that every tenth bed of flysch deposits can be traced over this distance. This allowed 196 synchronous regional markers to be defined among the Upper Cretaceous–Danian deposits of the Greater Caucasus. Their synchronism was determined based on the relative cyclite number and confirmed by paleontologic data. In other words, the Upper Cretaceous–Danian lithostratigraphic scale comprising over 400 stratigraphic units was elaborated. Its correspondence with the stratigraphic scale was confirmed by paleontologic evidence. The formation period of one global lithostratigraphic unit is expected to equal 22 ± 9 k.y., which corresponds to the formation time of one gilbert, or to the smallest of four Milankovich cycles. By 2059–2072, the standard Phanerozoic geochronologic scale with 18 000–46 000 stratigraphic units is expected to be eventually elaborated.

Reviewers Yu.B. Gladenkov and V.A. Krasilov

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Gondwana Glaciation and the Distribution of Corals during the Ordovician

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Abstract—The influence of the Gondwanaland glaciation on the distribution range of Late Ordovician corals is demonstrated. In the Late Ordovician, the Gondwana continent was affected by a considerable glaciation. Glacial deposits are known to exist in Saudi Arabia, Africa, South America, etc. However, no coral remains have been recorded in vast areas of marine Late Ordovician deposits contiguous to that part of Gondwanaland. Among areas with abundant Late Ordovician corals, the Canadian, Siberian, European, and Central Asian paleobiogeographic provinces and, apparently, the Australian region can be primarily mentioned. At the beginning of the Silurian period, the Late Ordovician geographical zones and biogeographical areas disappeared, whereas the geographical range of corals widened.

Key words: *Ordovician, Gondwanaland, glaciation, distribution of corals.*

INTRODUCTION

In the twentieth century, Paleozoic corals have been investigated in all the regions where they occur. It was found that the distribution of Ordovician corals was appreciably narrower with respect to Silurian, Devonian, and Late Paleozoic species. Localities of Late Ordovician corals are mostly concentrated in northern areas of the present northern hemisphere. They are also present in the United States, Central Asia, Kazakhstan, China, eastern Australia, and Tasmania. No coral remains were revealed in the Late Ordovician marine deposits of South America, Mexico, southern Europe, northern and southern Africa, Saudi Arabia, Turkey, Iran, Afghanistan, and certain regions adjacent to the western part of the former Gondwanaland.

During investigations of Ordovician corals, a large Late Ordovician glaciation of the western part of Gondwanaland was discovered.

GEOGRAPHICAL DISTRIBUTION OF LATE ORDOVICIAN CORALS

The Late Ordovician deposits contain the greatest number of the taxonomically most diverse corals. Fifteen orders and suborders, 40 families and subfamilies, and 160 genera were recorded in this interval. In stratified Ashgillian organogenic-detrital limestones, their remains occupy up to 10% (and more) of the volume of host rocks and dominate among all the fossils. However, the distribution range of Ordovician corals was appreciably smaller than that of their Silurian, Devonian, and Late Paleozoic species.

Within the present northern hemisphere, the Late Ordovician corals are widespread in the Baltic Region (43 genera), Scandinavia (41 genera), British Isles (30 genera), United States (42 genera except Alaska), Canada (55 genera), Greenland (15 genera), Alaska (20 genera), arctic Asia, including adjacent islands (65 genera), the Urals (30 genera), the Siberian platform (40 genera), Mongolia and Tuva (43 genera), Altai (52 genera), northeastern Kazakhstan (48 genera), southeastern Kazakhstan (44 genera), southern Tien Shan (45 genera), and northern (60 genera) and southern (40 genera) China. Thousands of specimens representing numerous genera have been collected from diverse localities. Corals of all the above-mentioned regions were encountered far away (more than 2000 km away) from the formerly ice-covered western part of Gondwanaland. Scarce Late Ordovician corals were recorded in Belgium (8 genera), Podolia (9 genera), the Pamirs (6 genera), and in the Himalayas (6 genera). No Ordovician corals were identified southward of the above-mentioned localities in America, Africa, Europe, and almost all of Asia. Only two small and intensely recrystallized tabulate fragments, presumably of *Quepora* sp., were described from the Malakka Peninsula by Thomas and Scrutton (1969). The graptolite dating of the host deposits suggested their Ordovician, probably middle Caradocian–early Ashgillian, age. The tabulates, being fragmentary, could not yield a more precise age than the Ordovician–Lludlovian. Within the southern hemisphere, Late Ordovician corals occur in eastern Australia (36 genera) and Tasmania (23 genera).

These are, virtually, all the regions that contain Late Ordovician corals. Silurian corals were also found in all the above-mentioned provinces. Moreover, Early Silurian corals were recognized (though no signs of Ordovician corals were observed) in the Antarctic

Table 1. The Late Ordovician paleoclimatic zoning

Paleoclimatic zones	Description	Regions
Proximal zone of the coral distribution range	Abundant and highly diverse corals. Numerous colonial rugose corals	Northern North America, Arctic region, Urals, northern Asia, northern China
Distal zone of the coral distribution range	Moderately diverse and abundant corals. Rugoses are mostly represented by solitary forms	United States, British Isles, Scandinavia, Baltic region, Central Asia, southern China, Australia
Şuperdistal zone of the coral distribution range	Occasional corals of a low diversity. Solitary rugoses only	Belgium, Podolia, Pamirs, Himalayas
Gondwana-related zone of the World Ocean	Brachiopods, trilobites, graptolites, conodonts. Corals are not recorded	Mexico, southern Europe, Near and Middle East, and some other areas
Western Gondwana	Glacial deposits	Eastern South America, Africa, Saudi Arabia, and other regions

region, South America, Mexico, southern Europe, Turkey, Iran, Afghanistan, and in some other places, while Devonian corals were recorded even in Africa. Such a restricted distribution of Ordovician corals is the result of a gigantic glaciation, which took place in the western part of Gondwanaland in the Late Ordovician. In the Late Ordovician, the South Pole was situated in Gondwanaland, approximately in the center of the Sahara (Havlicek, 1974).

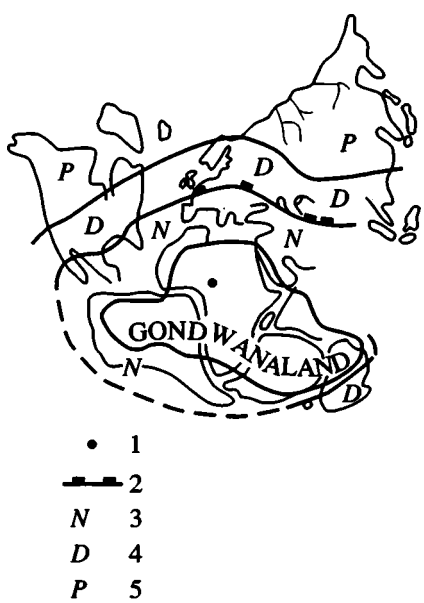
PALEOCLIMATIC ZONING IN THE LATE ORDOVICIAN

The Gondwana glaciation of the Late Ordovician time was responsible for the following paleoclimatic zoning (Table 1; figure):

(1) *Western Gondwana* was covered by a gigantic ice sheet. Late Ordovician glacial deposits are widespread in Saudi Arabia (McClure, 1988), the Sahara (Legrand, 1988), South Africa (Brenchley, 1988), Argentina, Bolivia, Brazil, Peru (Boucot, 1988), and in other regions. In the Late Ordovician, the Gondwana ice sheet was continent-wide, extending from the South Pole at least up to 40° S (Brenchley, 1988).

(2) *Gondwana coral-free zone of the World Ocean.* The Upper Ordovician deposits of this zone are abundant in South America (Argentina, Bolivia, Venezuela, and some others), Mexico, southern Europe (Czech Republic, Austria, Italy, France, Spain, Portugal), northern and southern Africa, Saudi Arabia, Turkey, Iran, Afghanistan, and in other regions. The Upper Ordovician marine deposits of this zone contain numerous brachiopods, mollusks, trilobites, graptolites, and conodonts, but corals have not been identified there. Ordovician glaciomarine deposits of this zone were distinguished in Normandy, Turingia, Spain, Portugal, and in other provinces (Brenchley, 1988). A zone of the horse-shoe shape surrounded the western part of Gondwanaland. Further east, it was narrower and less distinct, and in Australia and Tasmania it was absent.

(3) *The superdistal (marginal) zone of the distribution of Late Ordovician corals* was situated along the southern border of their habitat area. The zone was the narrowest and poor in corals. It is identified in Europe and in the western part of Asia. Tabulates and rugoses of this zone are found in Belgium (8 genera), Podolia (9 genera), the Pamirs (6 genera), and the Himalayas (6 genera). They are scarce and invariable. Tabulates are represented by small forms, and rugoses, by small and solitary species. Colonial rugoses have not been recognized in this zone. Sytova (1975) reported that 50 rugose corals were collected in the Late Ordovician of Podolia. They were represented by dwarf forms (10 to 30 mm high) of the *Kenophyllum, Leolasma, Neotryplasma,* and *Rectigrewingkia* genera. The Ordovician deposits in the Pamirs bear brachiopods, trilobites, ostracods, echinoderms, and other fossils, whereas corals are scarce here. Tabulates *Catenipora, Hemiagetolites, Palaeofavosites,* and small solitary rugoses *Grew-*



Paleoclimatic zoning of the Late Ordovician time: (1) the South Pole; (2) fragments of the superdistal (marginal) zone of the coral distribution range; (3) coral-free zone of the World Ocean; (4) distal zone of the coral distribution range; (5) the proximal zone of the coral distribution range.

ingkia and *Streptelasma* occur in the Late Ordovician beds of the northern Pamirs. They were all reported from one locality in the Peter the Great Range. In the Upper Ordovician of the central Pamirs, V.I. Dronov identified one tiny solitary rugose *Calostylis* (?) sp.

Corals of each region of the superdistal zone were closely associated with their counterparts confined to the nearly distal zone. Thus, all the genera of the Late Ordovician corals of Belgium are also recorded in the British Isles and Scandinavia; all genera of Podolia are recognized in the Baltic region; all genera occurring in the Pamirs, are identified in southern Tien Shan; and all genera common for the Himalayas are found in China.

(4) *The distal zone of coral distribution* is most likely located in southern Tien Shan (the Zeravshan–Gissar Mountains), southeastern Kazakhstan, the Baltic region, Scandinavia, the British Isles, and the United States (figure). Corals of southern China, Australia, and Tasmania can be also referred, though somewhat conventionally, to this zone.

Corals are much more diverse here, and their abundance is thousands of times higher than in the superdistal zone, though they are less diverse than in the proximal zone. Tabulates are dominated by Favositida, Halysitida, and Proporida. The cosmopolitan *Catenipora, Eofletcheria, Lyopora, Mesofavosites, Palaeofavosites, Propora*, and some other genera are widespread. Endemics common for the European and Central Asian provinces are also present (see below). Rugoses are dominated by solitary *Bodophyllum, Brachyelasma, Calostylis, Dalmanophyllum, Grewingkia, Streptelasma, Tryplasma*, and some other taxa. The diversity of colonial rugoses is significantly lesser with respect to solitary corals. Usually, no massive (cerioid) colonies are observed. For example, in the Late Ordovician of the Zaravshan–Gissar province, researchers have collected about 3000 solitary rugose specimens, 2000 tabulates, and only 15 specimens of small bushy (phaceloid) rugose colonies over 30 years, whereas no massive (cerioid) rugoses (Favistelidae and others) have been recorded in the Ordovician beds of Central Asia. No Ordovician Favistelidae have been identified in the Baltic region either (Kal'o *et al.*, 1970), though other corals are abundant there. Solitary rugoses prevail among Late Ordovician corals from the United States. There are plenty of representatives of *Streptelasma*. The diversity of tabulates (together with species of Geolithoidea) is second in significance. Colonial rugoses are rare and represented by phaceloids (Elias, 1991, 1992; Elias and Young, 1992).

(5) *The proximal (internal) zone* of the distribution of Late Ordovician corals was far away from Gondwanaland and was rich in corals. They were represented by all orders and suborders characteristic of the Late Ordovician. There are plenty of tabulates and rugose corals in the northern part of North America with adjoining islands, in Greenland and northern Asia with adjacent islands, and also in the Urals, Mongolia, Altai, northern China, and northeastern Kazakhstan. The proximal zone includes all regions of the Arctic paleozoogeographic province and some regions (northern China and northeastern Kazakhstan) of Central Asia.

Numerous cosmopolitan genera, such as *Calapoecia, Catenipora, Eofletcheria, Lyopora, Mesofavosites, Palaeofavosites, Plasmoporella, Brachyelasma, Grewingkia, Palaeophyllum, Streptelasma, Tryplasma,* and others occur in the proximal zone together with cyrtophyllids, tollinids, tetradiids, sibiriolitids, and favostellids, which are rare or absent in the distal zone and also were never known in the superdistal zone.

The above-described Ordovician zones were situated in the southern hemisphere and in the equatorial part of the northern hemisphere. In the Ordovician, the North Pole was in the central part of the Pacific Ocean. No Ordovician deposits were known to exist in this province and contiguous areas.

PALEOBIOGEOGRAPHICAL ZONES OF THE DISTRIBUTION OF LATE ORDOVICIAN CORALS

The above-described zones are not complete paleobiogeographical areas or provinces; they only correspond to the climatic belts that existed in the Late Ordovician. Usually, the paleogeographic area or province comprises regions with a direct faunal relationship and without appreciable difference in the habitat conditions of biota. One wide open marine basin could be situated within two, three, and even four zones, while one zone could pass through different marine reservoirs separated from one another. In the former case, the same paleobiogeographic province could lie within two or three zones, while in the latter, one zone could comprise several paleobiogeographic provinces.

Let us discuss the paleobiogeographic zoning in the distribution of Late Ordovician corals. Many researchers were involved in paleobiogeographic investigations of Ordovician corals, among them Sokolov (1951, 1962), Ivanovskii (1965), Kal'o, Klaamann, and Nestor (1970), Leleshus (1970), Lin Bao-yu and Webby (1989), Wang and Chen (1991), and others. Their paleogeographic schemes of the Late Ordovician, in addition to properly substantiated conclusions, comprise certain problematic issues, the decisions on which could change as new materials appear or the available data are reevaluated.

To establish paleobiogeographic zones, I used the statistical method. The taxonomic similarity between corals from different regions was found by the equation

$$R = \frac{C}{N_1 + 5F} \times 100, \quad (1)$$

where R is the degree of similarity; C is the number of common genera; N_1 is the number of genera of less frequently occurring taxa; and F is the number of families

Table 2. Taxonomic similarity between the Late Ordovician corals from different regions

Regions	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Baltic region	100	75	67	45	26	21	25	18	21	26	33	36	38	36	38
Scandinavia	75	100	80	48	30	23	24	17	20	27	33	33	40	35	46
British Isles	67	80	100	50	28	27	24	15	25	25	29	30	36	32	43
United States	45	48	50	100	62	40	35	30	31	31	34	33	28	30	50
Canada	26	30	28	62	100	76	51	68	54	50	38	20	15	20	41
Arctic region	21	23	27	40	76	100	60	70	58	54	46	22	20	25	34
The Urals	25	24	24	35	51	60	100	57	63	56	47	18	18	22	35
Siberian platform	18	17	15	30	68	70	57	100	70	50	30	14	13	15	26
Mongolia and Altai	21	20	25	31	54	58	63	70	100	67	60	34	20	23	47
Northern China	26	27	25	31	50	54	56	50	67	100	80	63	50	64	52
Northeastern Kazakhstan	33	33	29	34	38	46	47	30	60	80	100	77	53	53	50
Southeastern Kazakhstan	36	33	30	33	20	22	18	14	34	63	77	100	81	70	42
Southern Tien Shan	38	40	36	28	15	20	18	13	20	50	53	81	100	50	43
Southern China	36	35	32	30	20	25	22	15	23	64	53	70	50	100	56
Australia	38	46	43	50	41	34	35	26	47	52	50	42	43	56	100

widespread in one region and not known in the other. The number 5 was obtained empirically.

Let us estimate the degree of similarity between corals of the Baltic region and Siberian platform by equation (1). There are 43 genera in the Baltic Region and 40 ($N_1 = 40$) in the Siberian platform. Twelve of them are common ($C = 12$). Apart from this, among corals of the Baltic Region and Siberian platform, there are 5 families, representatives of which are abundant in one region and unknown in the other ($F = 5$). Families Cyrtophyllidae, Tollinidae, Sibiriolitidae, and Favistellidae are widespread in the Ordovician of the Siberian platform and unknown in the Ordovician of the Baltic region, whereas the Coccoseridídae family is widely distributed in the Baltic region and not recognized in the Siberian platform. Using these values in (1), we estimate the taxonomic similarity between corals of the Baltic region and Siberian platform.

$$R = \frac{12 \times 100}{40 + 25} = 18. \qquad (2)$$

Families or orders whose representatives are not recognized in one of the regions and scarce in another are not included in the equation. Rare and poorly studied genera, if they were recorded only in one locality of a region, are not taken into account either.

By equation (1) I estimated the degree of similarity between Late Ordovician corals of various regions. The results of my calculations are given in Table 2. This table is very informative. For instance, it shows that connections between corals of the Baltic region and the Urals developed through the fauna of Scandinavia, the British Isles, the United States, Canada, and the Arctic provinces. The degree of similarity is obvious for the European, Canadian–Siberian, and Central Asian palеozoogeographic zones, while the Australian province is less individual.

The European region (Western Europe and conventionally the United States, except Alaska). The Late Ordovician corals from the Baltic region, Scandinavia, British Isles, and other West European areas have close taxonomic similarities, and most researchers believe that they inhabited the same paleobiogeographic province. Seventy genera of Late Ordovician corals have been identified in Western Europe. Most common among them are tabulates, such as *Acidolites, Catenipora, Coccoserisa, Esthonia, Kierolites, Mesofavosites, Palaeofavosites, Palaeoporites, Protoheliolites, Sarcinula, Trochiscolithus,* and *Wormsipora*, and also rugose corals *Braehyelasma, Calostylis, Grewingkia, Palaeophyllum, Streptelasma,* and *Tryplasma*. Nearly all the above-mentioned genera occur in the United States. Many tabulates, rugoses, brachiopods, and other invertebrate fossils were common in both Western Europe and the United States during the Late Ordovician. In addition, there are many identical genera and species among the Ordovician corals of the United States and Canada. The Ordovician corals of Canada are most similar to those of the Arctic region and inhabited the Arctic area. Thus, the Late Ordovician corals of the United States represented an intermediate fauna between corals of Western Europe and Canada. Late Ordovician corals of Western Europe are taxonomically similar to those of the United States and, to a lesser degree, to corals of the Central Asian and Australian regions, though they greatly differ from corals of the Arctic region. Sokolov (1951) was the first to note a great taxonomic difference between the Late Ordovician corals of Western Europe and those of Arctic areas.

My conclusions are different from those of Wang and Chen (1991), who suggested that faunas of northern Europe, southeastern Asia, and Australia belong to the eastern Prototethys, whereas faunas of North America, Siberia, Kazakhstan, and northern China are referred to the western areas of this paleoocean.

Canadian–Siberian region (northern North America, Greenland, Arctic areas, Urals, northern Asia, Altai, and Mongolia). One hundred twenty genera of Late Ordovician corals have been identified in the Arctic region. Most common are tabulates *Calapoesia, Catenipora, Cyrtophyllum, Eofletcheria, Foerstephyllum, Karagemia, Nyctopora, Palaeofavosites, Paratetradium, Phytopsis, Rhabdotetradium, Rhaphidophyllum, Saffordophyllum, Sibiriolites, Tetradium, Tollina,* and *Vacuopora* associated with rugoses *Bighornia, Deiracorallum, Favistella, Grewingkia, Kenophyllum, Lobocorallum, Palaeophyllum*, and some other taxa.

The Arctic region discussed here is correlative with the Canadian–Arctic and Siberian regions that were studied by Sokolov (1962). Works by Sokolov provided much information on the geographical distribution of Late Ordovician corals. These data revealed a great taxonomic similarity between the Late Ordovician corals of Canada, the Arctic region, Siberia, and northeastern Russia. Wang and Chen (1991) considered the Arctic region as the western Prototethys and assigned the Late Ordovocian fauna of the United States, Kazakhstan, and northern China to this kingdom. The evidence we discuss suggests that Late Ordovician faunas of Kazakhstan and northern China are closer to those of Central Asia than to the Arctic faunas, whereas the latter greatly differ from coral assemblages of Central Asia.

Central Asian region (Central Asia, Kazakhstan, and China). About 100 genera of Late Ordovician corals have been identified in this region. Most common are tabulates *Acdalopora, Agetolitella, Agetolites, Catenipora, Granulina, Halysites, Hemiagetolites, Lyopora, Mesofavosites, Palaeofavosites, Plasmoporella, Propora, Reuschia, Brachyelasma, Calostylis, Grewingkia, Helicelasma, Palaeophyllum, Pseudophaulactis, Streptelasma,* and *Tryplasma*. Corals occurring in northern China and northeastern Kazakhstan are intermediate in appearance between corals of the Arctic and the Central Asian provinces. Though there were broad and direct relations between faunas of the Arctic and Central Asian provinces through northern China, Altai, and some other regions, corals of the Central Asian province are, nevertheless, more similar to their assemblages from remote Western Europe than to those from the neighboring Arctic region. This might be a consequence of the above-described zoning. Corals of Western Europe, southern Kazakhstan, Central Asia, and southern China inhabited the distal zone of their distribution range, whereas corals of the Arctic region were concentrated in the proximal zone. Wang and Chen (1991) did not identify the Central Asian paleozoogeographical province. They attributed the northern part of the region, including corals of Siberia, the Arctic region, and North America, to the western Prototethys, while the southern coral communities of Western Europe and Australia were considered to belong to the eastern Prototethys.

Australian region (Australia and Tasmania). Forty-two genera of Late Ordovician corals have been recorded in this region. Their endemic features are not distinct. All the families and almost all genera (except *Bowanophyllum* and *Hillophyllum*) are also encountered in other paleozoogeographical areas. There are genera typical of Europe (*Acidolites, Coccoseris, Reuschia, Grewingkia, Streptelasma, Tryplasma*, and others), the Arctic region (*Bajgolia, Calapoecia, Cyrtophyllum, Cyathophylloides, Foerstephyllum*, and others), and the Central Asian paleogeographic province (*Agetolites, Granulina, Lyopora, Plasmoporella, Grewingkia, Streptelasma*, and others). Table 2 indicates that Late Ordovician corals of Australia have a close taxonomic similarity with corals of southern China and greatly differ from corals of the Arctic region. A close taxonomic relationship between the Ordovician fauna of Australia and that of southeastern Asia has been emphasized by many researchers (Burrett and Stait, 1985; Lin Baoyu and Webby, 1989; and some others).

At the beginning of the Silurian, the giant ice sheet of Gondwanaland disappeared, and a global transgression took place. The Late Ordovician paleogeographical zones and provinces along with distribution ranges of biota changed. The distribution area of Llandoverian rugose corals and tabulates was much wider than that of the Ordovician time, and they were represented by cosmopolitan genera and species. The Early Silurian coral communities of Afghanistan, Iran, Central Asia, Kazakhstan, China, Siberia, the Arctic region, western Europe, Greenland, and North America were dominated by common genera and by many common species.

The Late Ordovician glaciation of Gondwanaland is well confirmed by a complete absence of corals in the wide equatorial belt (1000 to 2000 km wide), contiguous to the western part of the paleocontinent. Silurian, Devonian, Carboniferous, and Permian corals dwelt in the close vicinity of western and eastern Gondwanaland. Their assemblages are abundant in South America, southern Europe, Near and Middle East (Turkey, Iran, Afghanistan, and some other areas), Indochina, and Australia. Devonian, Carboniferous, and Permian corals were also revealed in Africa.

The paleogeographic changes at the end of the Ordovician–beginning of the Silurian greatly affected the evolution of corals. About 70% of the Ashgillian coral genera disappeared at the end of the Ordovician. All rare and endemic genera, along with certain genera intermediate between endemic and cosmopolitan, such as *Agetolites, Agetolitella, Granulina, Hemiagetolites, Plasmoporella, Favistella*, and some others, became extinct. Most of the Late Ordovician cosmopolitan gen-

era continued their evolution in the Silurian time. A new intensive speciation of corals started in the Llandoverian. Nearly 100 genera appeared at that time, among them *Favosites, Entelophyllum*, and some others.

CONCLUSION

(1) In the Ordovician, the geographical distribution of corals was narrower than in later Paleozoic periods. Late Ordovician corals were mainly concentrated in northern areas of the present northern hemisphere. No coral remains were found in the Ordovician marine deposits of South America, Mexico, southern Europe, Africa, Saudi Arabia, Near and Middle East, or other regions confined to the western part of the former Gondwanaland.

(2) The Late Ordovician paleoclimatic zoning from the south to the north resulted from the Gondwana glaciation and was as follows (figure):

(a) The ice-covered western part of Gondwanaland;

(b) The coral-free zone of the World Ocean adjacent to Gondwanaland;

(c) The superdistal (marginal) zone of the coral distribution range, where corals were rare and of a low diversity;

(d) The distal zone of the coral distribution range with numerous tabulates and solitary rugose corals;

(e) The proximal zone of the coral distribution range inhabited by abundant and diverse corals.

(3) In the Late Ordovician, the Canadian–Siberian, European, and Central Asian paleogeographic provinces and less clearly the Australian province were distinguished by corals.

(4) At the beginning of the Silurian, the giant ice sheet disappeared, and the geographical distribution of corals became wider. The former paleoclimatic zones and paleobiogeographic provinces vanished.

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Comments on the General Stratigraphic Scale of the Quaternary System in Russia

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One of the principal goals of Quaternary investigations is to elaborate the Standard Stratigraphic Scale for the latest geological period in the earth's history. This goal is important from the viewpoint of existing theoretical concepts and in the practical aspect of geological mapping. This problem has been studied by many specialists in Quaternary and Neogene geology and has been a subject of discussion at numerous international congresses, symposia, and meetings.

The comprehensive investigations performed by Commissions on Stratigraphy of the International Union of Geological Sciences (IUGS) and on Stratigraphy of the International Union for Quaternary Researchers (INQUA) resulted in the approval of the Neogene–Quaternary boundary position at the chronological level of 1.6 Ma (International Geological Congress, 1984). The boundary stratotype section in southern Italy has been thoroughly studied (Cowie and Bassett, 1989; Aguirre and Pasini, 1985). In Russia, the Neogene-Quaternary boundary was placed long ago at the level of 0.7 Ma, although other alternatives (1.6, 1.8, and 2.4 Ma) were also suggested. Because of a different understanding of the same stratigraphic units in Russia and other countries, it appeared necessary to modify the general stratigraphic scale of Quaternary deposits in Russia. In 1991, the Interdepartmental Stratigraphic Committee (ISC) of Russia approved a scale that envisaged the position of the lower Quaternary boundary at the level of 1.6 Ma and the division of the Quaternary System into three series: Eopleistocene (1.6–0.8 Ma), Pleistocene (0.8–0.01 Ma), and Holocene (younger than 0.01 Ma). In 1994, the All-Russian conference on Quaternary research discussed the problem of Quaternary subdivisions and recommended the ISC to proceed with the relevant investigations. Further studies resulted in the construction of the general Quaternary stratigraphic scale (Table 1), which incorporated the smaller taxonomic units used in Russia and also well grounded their correlation with the Quaternary part of the Global Stratigraphic Scale (Cowie and Bassett, 1989). Although it was published in the abstract volume, this scale was not discussed at the INQUA Congress in 1995.

In February 1995, the ISC decided to include the Eopleistocene in the Pleistocene and to define the range of the latter as it was accepted by the International Commission on Stratigraphy. However, this raised the question on how to divide such an extremely extended Pleistocene.

As is known, the division of the Pleistocene into the lower (1.6–0.8 Ma), middle (0.8–0.13 Ma), and upper (0.13–0.01 Ma) parts is a common practice in western Europe (see Table 1). However, many Russian specialists do not find this procedure adequate because these three subdivisions greatly differ from each other both in duration (0.12 and 0.8 m.y.) and the number of related events. It seems to be more appropriate to divide the "great" Pleistocene into two nearly equal parts corresponding to the Eopleistocene (Nikiforova and Aleksandrova, 1991) and the Pleistocene, as it was decided long ago in Russia. This two-member structure of the Pleistocene was adopted by the ISC in 1995 (see Table 1). The name "Eopleistocene" was retained for the subdivision corresponding to the lower Pleistocene of the West European scale (1.6–0.8 Ma). The subdivision corresponding to the interval of 0.8–0.01 Ma was named Neopleistocene and subdivided into the lower (0.8–0.4 Ma), middle (0.4–0.13 Ma), and upper (0.13–0.01 Ma) parts (links). The upper link corresponds to the upper Pleistocene, and the lower and middle links together, to the middle Pleistocene of the European scale. In this case, the ISC scale would retain all the boundaries recognized in western Europe and all Quaternary subdivisions (under slightly altered names) used in Russia (the continuity principle is known to be the most important in creating any stratigraphic scale). Authors of the scale believe that it has some advantages over the West Euro-

Table 1. Correlation between the Quaternary nomenclatural units adopted by the Interdepartmental Stratigraphic Committee of Russia and recommended by the Commission on Stratigraphy of the International Union of Geological Sciences (IUGS)

Scale of the Interdepartmental Stratigraphic Committee of Russia, 1995						Scale approved by IUGS				
system	series	subseries	link	step	major chronological boundaries, (Ma)	system	series	stage	major chronological boundaries, (Ma)	magneto-stratigraphy
Quaternary	Holocene				0.01	Quaternary	Holocene		0.01	Brunhes (Magnetochron 1)
	Pleistocene	Neopleistocene	upper	Fourth			Pleistocene	upper		
				Third						
				Second						
				First						
			middle					middle		
			lower		0.8					
		Eopleistocene	upper					lower		Matuyama (Magnetochron 2)
			middle		1.6				1.6	
Neogene	Pliocene	upper				Neogene	Pliocene	upper		Gauss (3)

pean scale, because this approach excludes the coupling of subdivisions corresponding to different stages of faunal and floral development, such as Cromerian–Elsterian and Holsteinian–Saalian (even in western Europe, the lower–middle and middle–upper Pleistocene are in frequent use as independent units).

Authors of the scale also consider it expedient to accept steps (recommended in the Russian *Stratigraphic Code*) as the most detailed units of the general scale corresponding to O-isotope stages in oceanic sediments, although the latter cannot be unambiguously correlated with continental deposits.

However, some other Russian specialists argue that the former division of the Quaternary System into the upper, middle, and lower Pleistocene and Eopleistocene (corresponding to the older interval of 1.8–0.7 Ma) has been justified by investigations of continental and marine sequences in the vast territory of northern Eurasia during many years. Such a subdivision of the Quaternary System has purely nomenclatural differences from that adopted by the ISC in 1995 and corresponds to large biostratigraphic subdivisions based on changes in faunal assemblages (Gromov, 1948) and phylogenetic successions of paleoflora (Grichuk, 1973).

Authors of the general stratigraphic scale of the Quaternary System approved by the ISC see its main advantage in its correspondence with the scale approved by the commissions on stratigraphy of INQUA and IUGS (Cowie and Bassett, 1989) and popular in other countries. The general scale was discussed at two sessions of the Commission for Quaternary Research, Russian Academy of Sciences, which were attended by several members of the International Subcommission on Neogene Stratigraphy of IUGS. Some critical comments were made at these sessions. For instance, some of us (A.A. Velichko and Yu.A. Lavrushin) indicated the decorative character of the accepted alterations, because one of the accepted nomenclatural units (Eopleistocene) is of the subseries rank, whereas subdivisions of the second half of the system (lower, middle, and upper Pleistocene) are ranked as links. In addition,

Table 2. A version of the general Quaternary stratigraphic scheme suggested by A.A. Velichko and Yu.A. Lavrushin

System	Series	Subseries	Stage
Quaternary	Holocene		
	0.01		
	Pleistocene	Neopleistocene	Upper Pleistocene
		0.125	
		Plenipleistocene	Middle Pleistocene
			Lower Pleistocene
	0.8		
	Eopleistocene		Upper Eopleistocene
			Lower Eopleistocene
1.6			
Neogene	Pliocene		

the ISC scale unites the three last units under the name of "Neopleistocene" (the subseries that succeeds Eopleistocene). When accepted, this new term makes it possible to apply the name "Pleistocene" to the entire Quaternary System (excluding the Holocene), as in other national scales. The opponents also foresee problems regarding the correlation of different scales after the introduction of the new term "Neopleistocene." This term is to be used internationally, which would certainly create problems in the correlation procedure for researchers in some countries.

Moreover, it is difficult to sustain the common status of the "Neopleistocene" subdivisions relative to that of the "Pleistocene" subdivisions of other national scales. For instance, the upper Pleistocene of foreign scales is identical to this unit of the national scale in terms of chronostratigraphy and structure, whereas the middle and lower Pleistocene are uncommensurable in their scope with the corresponding units of the general Quaternary scale of Russia. It should also be remembered that the term "Neopleistocene" had been already suggested for the interval of the upper Pleistocene when the term Eopleistocene was applied to the unit of the different stratigraphic range.

Criticism of the scale adopted by the ISC shows that a new approach to the correlation of different stratigraphic scales of the Quaternary System is necessary. One possible version is presented in Table 2. The Quaternary subdivisions suggested in the table are ranked as follows: the largest units, such as the Eopleistocene and Pleistocene, along with the smaller Holocene are accepted as series; the Plenipleistocene and Neopleistocene, as subseries, and the lower and upper parts of Eopleistocene and the lower, middle, and upper parts of the Pleistocene, as stages.[1]

The adopted taxonomic subdivisions of the Quaternary System do not principally contradict *Stratigraphic Code,* are successive with respect to stratigraphic units accepted in the former Soviet Union, and can be easily correlated with those of the Standard Stratigraphic Scale approved by the International Commission on Stratigraphy. However, the introduction of such units as step and link into the general stratigraphic scale of the Quaternary System is a subject of controversy. Taking into consideration principles of their recognition and specific features of the Quaternary Period, it was suggested to consider them as regional units, i.e., the units of stratigraphic scales of an individual continent or ocean. In this connection, the idea was put forward to elaborate provincial stratigraphic scales for the large countries or their groups and for individual large sedimentary basins of recent oceans or inner seas. Finally, the local stratigraphic scales of the lowest stratigraphic rank may fit the territories of small countries or individual regions of large countries. Thus, some researchers (in particular, A.A. Velichko and Yu.A. Lavrushin) suggest a principally different approach to the elaboration of the Global Stratigraphic Scale of the Quaternary System.

Quaternary deposits in the vast territory of Russia have been intensively studied for a long time. Long research traditions, continuous renewal of concepts, utilization of new data on different provinces, and introduction of new methods into practice should serve as a basis for developing new stratigraphic scales of local and subglobal significance.

Therefore, the ISC scale is proposed for comprehensive discussion.

We believe that, in spite of possible criticism, this scale can be submitted to the Commission on Stratigraphy of INQUA for an international discussion to further refine the Global Stratigraphic Scale of the Quaternary System.

Reviewer L.A. Nevesskaya

[1] Some of us are against this suggestion because any stage must have its own name and stratotype.

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First Recognized Paleoeuxinian–Uzunlarian Deposits on the Bulgarian Coast

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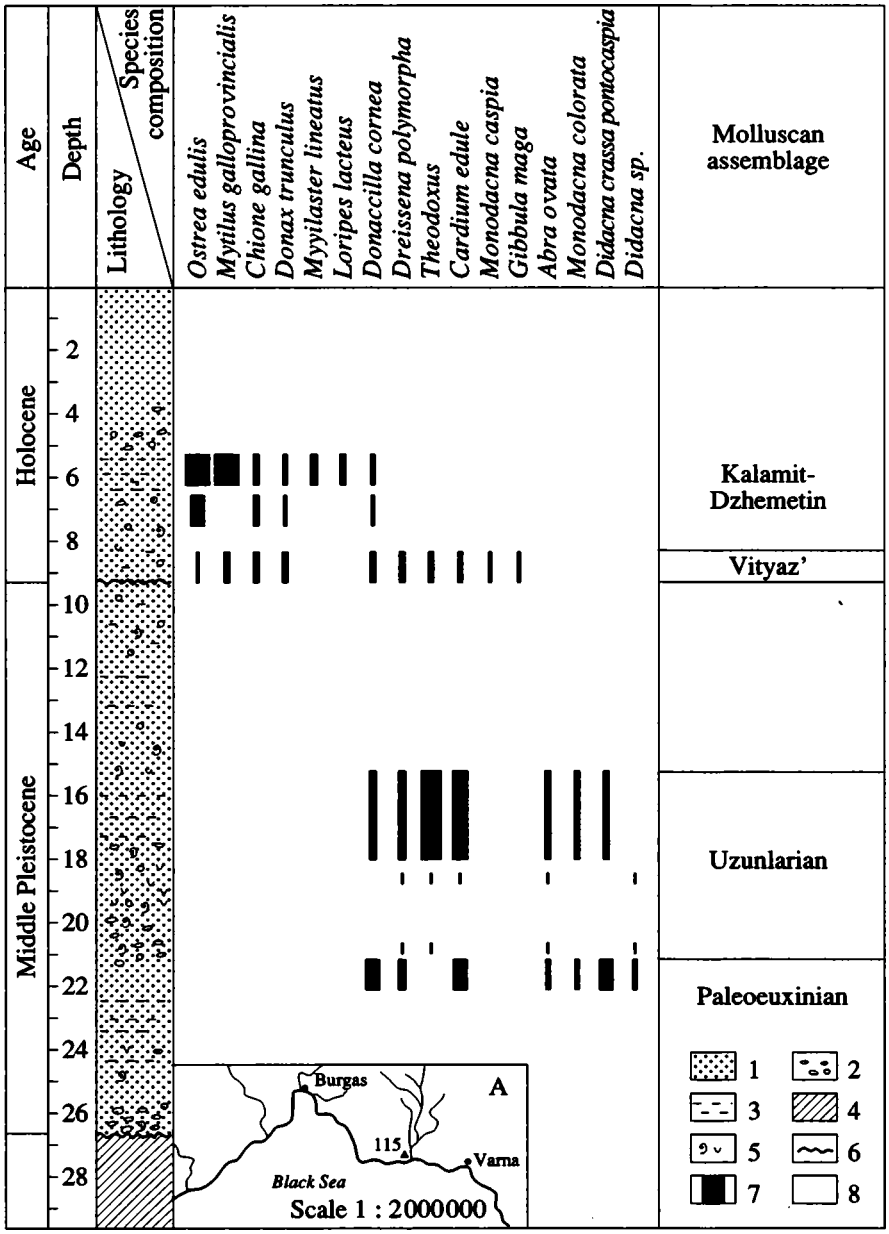
Paleoeuxinian and Uzunlarian deposits were recognized by Arkhangel'skii and Strakhov (1932a) at the Crimea–Caucasian coast of the Black Sea and later were described by many researchers (Muratov, 1960; Fedorov, 1963; Mikhailesku *et al.*, 1991; and others). On the northern coast of the Black Sea, these deposits form a sedimentary cover of marine terraces, the stratigraphical structure of which has been carefully studied. Up to the present, the Uzunlarian and Paleoeuxinian sediments bearing fossil fauna have not been known to occur on the Bulgarian coast. According to some researchers (Popov and Mishev, 1974; Fedorov *et al.*, 1962), there are marine terraces 35–65 m high formed in these periods. Surface beds of these terraces contain poorly preserved valves of *Ostrea,* and *Mytilaster lineatus*. However, because these fossils were not found in situ and do not belong to the key Uzunlarian and Paleoeuxinian species, the age determinations, which are based on the geomorphological position of the terraces and on the analogy with the Caucasian coast but not on biostratigraphic records, should be considered conventional.

The most complete Pleistocene sequences of the Bulgarian coast were recovered by several boreholes in the mouth areas of the Camchia and Fyndykliiskaya river valleys. The fossiliferous Paleoeuxinian and Uzunlarian marine deposits were found for the first time on the Bulgarian coast at the depth interval of 9–33 m (the Camchia boreholes 115, 116, and 117). Borehole 115 penetrated the most complete sequence. At the depth interval of 9.0–16.8 m, beneath the sand–gravel–pebble sediments (figure) with Holocene marine mollusks, there is gray and brown–gray compact silt with ferrugination spots, thin interbeds of fine-grained sand, and inclusions of carbonates, gravel, and rare pebbles. The deposits contain abundant casts and less frequent complete shells of mollusks, mainly of *Cardium edule lamarcki, Monodacna caspia*, and *Dreissena polymorpha.* Species *Gibbula maga, Abra ovata, Didacna crassa pontocaspia, Didacna* sp., *Monodacna colorata,* and *Theodoxus pallasi* were also found here. The shells are especially abundant in the lower part of these deposits grading into the underlying bed. The 16.8- to 21.3-m depth interval consists of light gray and gray compact silt alternating with sand and yellow fine- to small-grained loose sandstone with ferrugination spots, small pebbles, and gravel of dark-colored rocks. The sandstone contains fragmented and complete valves of large thick-walled shells of *Didacna crassa pontocaspia*, abundant *Dreissena polymorpha, Monodacna caspia*, and sporadic *Didacna* sp., *Monodacna colorata,* and *Abra ovata*. At the depth interval of 21.3–25.5 m, there is sand and yellow–gray fine- to small-grained loose sandstone with gray silt interbeds. Poorly preserved shells of mollusks, including *Didacna* sp., *Monodacna caspia, Dreissena polymorpha,* and *Theodoxus pallasi*, were found there. These deposits grade into the basal bed (25.5–26.3 m) of sand and brownish gray loose, poorly sorted sandstone of variable grain size enclosing abundant gravel and pebbles. This bed lies on poorly sorted and eroded continental deposits (26.8–34.0 m), which are conventionally dated by their position back to the late Pliocene–Pleistocene. Down to the bottom of the borehole, there are various marine and alluvial-marine Neogene deposits.

In the lithological aspect, the sediments from the depth interval of 9.0–26.8 m represent a single stratum containing two molluscan assemblages (see figure). The lower assemblage is dominated by brackish-water Caspian and Asov–Caspian elements (*Didacnacrassa pontocaspia, Monodacna caspia,* and *Dreissena polymorpha*). The Mediterranean fauna (single finds of *Abra ovata*) appears in the upper part of the beds enclosing the assemblage. These facts, as well as the predominance of the guide Paleoeuxinian species *Didacna crassa pontocaspia,* allow us to refer with confidence the enclosing sediments to the Paleoeuxinian basin. The upper assemblage is mainly represented by Mediterranean species (with predominant *Cardium edule lamarcki*) associated with a significant amount of Asov–Caspian elements (*Monodacna caspia* and *Dreissena polymorpha*). The continuous deposition of sed-



Deposits and molluscan assemblages (Borehole 115). (1) Sand; (2) pebble and gravel; (3) silt; (4) continental loams; (5) molluscan shells; (6) erosion surface; (7) abundance of shells (the interval thickness denotes the relative frequency: predominant, numerous, rare, single); (8) location of boreholes on the Black Sea coast in Bulgaria.

iments with the molluscan assemblages and a relatively gradual replacement of one (predominantly Caspian) fauna by another (predominantly Mediterranean) indicate the Uzunlarian age of the upper molluscan assemblage and enclosing deposits. It is noteworthy that similar but impoverished molluscan assemblages were recorded from the Uzunlarian–Paleoeuxinian deposits recovered by Borehole C-8 in the Burgas Bay shelf (Khrischev and Shopov, 1979). The upper assemblage is represented here by the association of *Dreissena polymorpha–Monodacna subcolorata–Abra ovata,* and the lower one, by the association of *Dreissena polymorpha–Didacna* sp.

Sequences recovered by the adjacent boreholes of the profile have similar structures and faunal assemblages. They also demonstrate the absence of the Karangatian deposits in the valley floor, which was probably caused by fluvial erosion in the Neoeuxinian time and by abrasion during the Holocene transgression. The Karangatian marine sediments are likely to be preserved only along the valley slopes. We identified these deposits containing the monodominant (*Corbula*

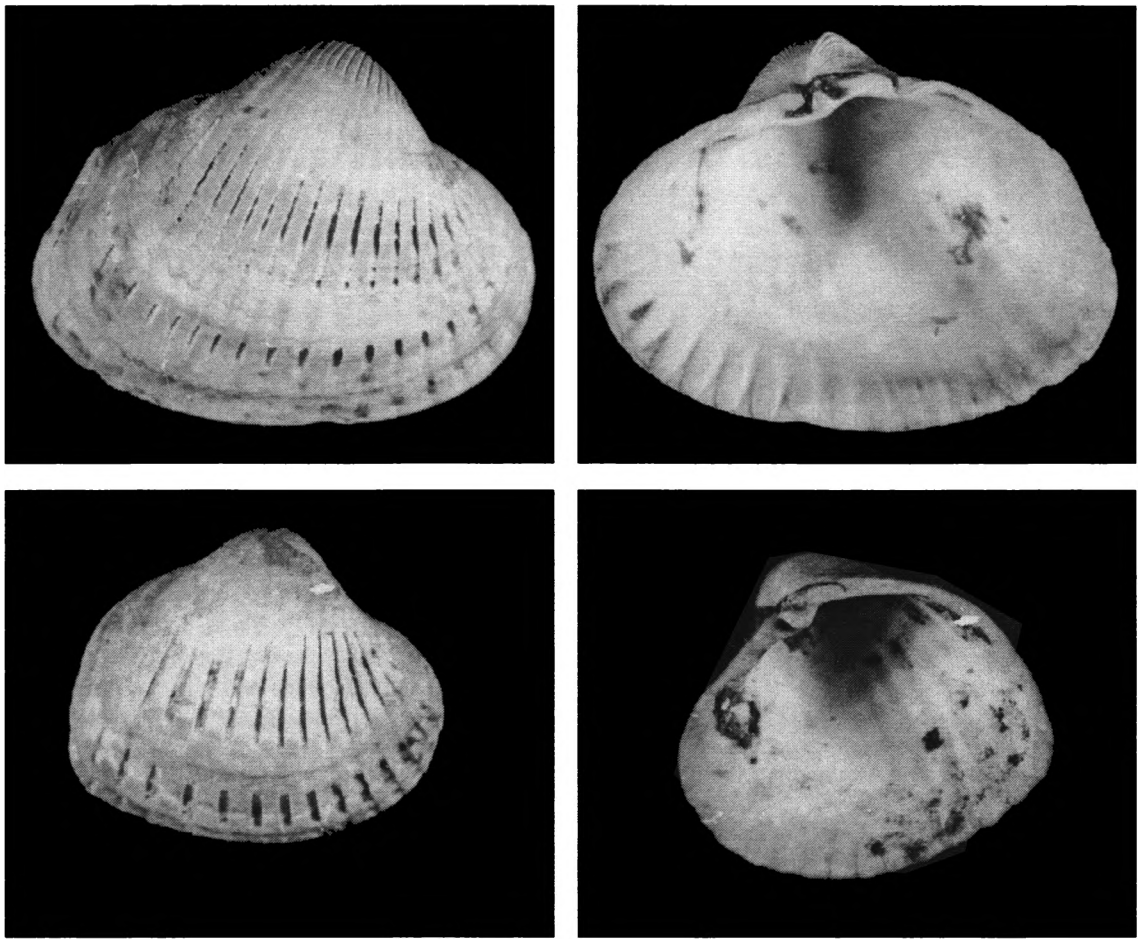


Plate. *Didacna crassa pontocaspia,* the inner and outer surfaces of shells, right valves. Borehole 115, the mouth area of the Camchia River valley, depth interval of 16.8–1.3 m.

gibba) molluscan assemblage in the terracelike scarp in the mouth part of the Fyndykliiskaya River valley.

The data presented above suggest that, in the middle Pleistocene time, the water masses of the Paleoeuxinian and subsequent Uzunlarian seas penetrated along the river valleys deep into the Bulgarian coast of the Black Sea. These were marine basins with water similar regimes near the Crimea–Caucasian and Bulgarian coasts. At first, the structure represented a closed basin of the Caspian type, which was afterwards filled with Mediterranean waters. Therefore, it is reasonable to suppose that the first stage corresponded to the low stand of the global sea level in the Paleoeuxinian, which was then replaced by the high sea-level stand during the Uzunlarian time.

Taking into consideration that *Didacna crassa pontocaspia* from the borehole sequences is the reference species for the Paleoeuxinian deposits now recognized for the first time on the Bulgarian coast and shelf, its brief systematic description and representation (Plate) are given below. The species *Didacna crassa pontocaspia* (Pavl.) was first identified by Pavlov (1925) as *Cardium pontocaspicum* from the lower horizon of Babel. Then Arkhangel'skii and Strakhov (1932a, 1932b) published descriptions of *Didacna pontocaspia* var. *elongata* Pavl. and *D. pontocaspia* var. *media* Pavl., which they found in the Paleoeuxinian deposits of the Caucasian coast and the Danube delta. Nevesskaya (1963) presented a detailed description of the species *D. crassa pontocaspia* from the Paleoeuxinian beds of the Kerch and Taman' peninsulas, the Azov Sea coasts, and the Caucasus.

Genus *Didacna* Eichwald, 1938

Didacna crassa (Eichwald, 1829)

Didacna crassa pontocaspia (Pavlov, 1925)

The shell is middle-sized (length 32–43 mm, height 27–33 mm), trigonally suboval, high (the height/length ratio is 0.77–0.80), convex (the convexity/height ratio, 0.40–0.45; the convexity/length ratio, 0.30–0.37), and slightly inequilateral (the ratio of the anterior area length to the general length is 0.46–0.47). Its beak extends far over the hinge area (the ratio of the beak height to the general height is 0.15–0.11). The beak is carinate, recurved, and usually displaced forward,

though it is sometimes central. The keel fold is distinct at the beak and flattened below. There are 25–31 ribs in total: 18–22 in the anterior area and 7–9 in the posterior area. The hinge is typical of this kind of species. The inner surface is smooth but becomes costate below the pallial line.

The description shows that the shells of *Didacna crassa pontocaspia* from the Bulgarian coast do not differ in morphology from those described by Pavlov (1925) and Nevesskaya (1963) from the Caucasian coast of the Black Sea.

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